Global Observations and Understanding of the General Circulation of the Oceans

Proceedings of a Workshop

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Global Observations and Understanding of the General Circulation of the Oceans

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August 8-12, 1983
Woods Hole Study Center

Ocean Climate Research Committee
Board on Ocean Science and Policy
Commission on Physical Sciences,
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The workshop was organized to address: (1) To determine whether we do indeed have the ability to obtain ocean data on a global scale that could profoundly change our understanding of the circulation; (2) to identify the primary and secondary elements needed to conduct a World Ocean Circulation Experiment (WOCE); (3) if the ability is achievable, to determine what the U.S. role in such an experiment should be; and (4) to outline the steps necessary to assure that an appropriate program is conducted. Background papers prepared by participants for background material at the workshop are included in the appendix to the proceedings.
The workshop was organized into three working groups: "Water Masses and Their Exchange", Atmosphere-Ocean Exchange", and Velocity Field and their reports are included in the proceedings.
Consensus of the workshop was that a World Ocean Circulation Experiment appears feasible, worthwhile, and timely. The workshop participants did agree that such a program should have the following overall goal: To understand the general circulation of the global ocean well enough to be able to predict ocean response and feedback to long-term changes in the atmosphere. Discussion of the overall goal, specific objectives, and recommendations for next steps in planning such an experiment are included in the text.

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Board on Ocean Science and Policy
National Research Council
2101 Constitution Avenue, N.W.
Washington, D.C. 20418

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PREFACE

A number of technical developments have the potential for greatly improving the observational data base available to oceanographers studying the general circulation of the ocean. This potential improvement bears directly on the problems in oceanography that involve large scales and long time periods and that have a direct impact on society. Societal issues involved include the growing concern about the role of the ocean in climate and climate change, the response of the ocean to increasing carbon dioxide in the atmosphere, and the long-term behavior of radioactive wastes deposited in the sea. The new technology holds much promise and is already available in principle, but questions have been raised as to whether the new observational systems based on this technology can or will be used to address the many issues now facing oceanographers.

The technical developments and recent advances in understanding of relevant ocean processes suggested to a number of interested scientists that a structured discussion of the problems of making global observations of the ocean would be particularly timely. In response, the Ocean Climate Research Committee of the Board on Ocean Science and Policy of the National Research Council held a workshop on "Global Observations and Understanding of the General Circulation of the Oceans." This report presents the proceedings of that workshop, which was held August 8 through 12, 1983, at the Woods Hole Study Center of the National Academy of Sciences, Woods Hole, Massachusetts.

A Steering Committee was appointed to organize the workshop, and they identified several issues to address: (1) To determine whether we do indeed have the ability to obtain ocean data on a global scale that could profoundly change our understanding of the circulation; (2) to identify the primary and secondary elements needed to conduct a World Ocean Circulation Experiment (WOCE); (3) if the ability is achievable, to determine what the U.S. role in such an experiment should be; and (4) to outline the steps necessary to assure that an appropriate program is conducted.

Fifty-eight individuals, representing physical, chemical, and biological oceanography, were invited to participate in the workshop (see Appendixes I and II). Before the workshop, some participants prepared papers on the general concepts of global observations. The compiled background papers were printed and bound prior to the workshop, where they served as a basis for discussion. The papers cover an

enormous range of issues, from specifics of particular instrument accuracies to semicomplete outlines of concepts of the entire World Ocean Circulation Experiment (see Appendix III). These papers represent personal views, and it was recognized that not all the recommendations and suggestions could be carried out.

Three working groups were organized during the workshop to consider specific technical issues and present their findings at plenary meetings:

- GROUP 1, "Water Masses and Their Exchange": William J. Jenkins and Joseph L. Reid, Co-Chairmen; Bert R.I. Bolin, Neil L. Brown, Kirk Bryan, Curtis A. Collins, James Crease, Robert Etkins, Manuel E. Fiadeiro, Joseph Pedlosky, Peter B. Rhines, Claes G.H. Rooth, Jorge L. Sarmiento, John Steele, Henry Stommel, Bruce Warren, and Ray F. Weiss.
- GROUP 2, "Atmosphere-Ocean Exchange": Pearn P. Niiler, and William George Large, Co-Chairmen; Joost A. Businger, Anthony Calio, D.E. Harrison, Robert H. Heinmiller, Jr., A.D. Kirwan, Ryuji Kimura, John Morrison, William C. Patzert, James F. Price, Roger Revelle, Robert H. Stewart, R.W. Stewart, and John Woods.
- GROUP 3, "Velocity Field": Harry L. Bryden and Walter H. Munk, Co-Chairmen: Robert E. Cheney, Russ E. Davis, Michael Ghil, Dale E. Haidvogel, Michel Pierre Lefebvre, James R. Luyten, James G. Marsh, Paola M. Rizzoli, Allan R. Robinson, Robert C. Spindel, George Veronis, Douglas C. Webb, and Stan Wilson.

Each working group prepared a report of its findings.

In organizing the workshop, the Steering Committee attempted to assure that a wide spectrum of views would be represented. The result was often heated and wide-ranging debate; nonetheless, by the end of the workshop, consensus did develop that a World Ocean Circulation Experiment appears feasible, worthwhile, and timely. The workshop participants made no attempt to design such an experiment, but did agree that such a program should have the following overall goal: <u>To understand the general circulation of the global ocean well enough to be able to predict ocean response and feedback to long-term changes in the atmosphere</u>.

Discussion of the overall goal, specific objectives, and recommendations for next steps in planning such an experiment are included in the text.

Financial support for the workshop was provided by the National Science Foundation, the National Oceanic and Atmospheric Administration, and the National Aeronautics and Space Administration.

Carl I. Wunsch,
Steering Committee Chairman
D. James Baker
Francis P. Bretherton
Wallace Broecker
James C. McWilliams
Worth D. Nowlin, Jr.
Ferris Webster

CHAPTER 1

INTRODUCTION

Lack of adequate knowledge of the general circulation of the ocean is a fundamental obstacle to solving many important problems of climate, ocean chemistry, and biology. The ocean is a global turbulent fluid, with temporal and spatial variability on all scales. There is agreement that the observations necessary to define all the important time and space scales everywhere are lacking. The main difficulty is our present inability to observe the three-dimensional ocean over the requisite long time scales.

It is possible that recent technical developments could significantly improve the observational base available to oceanographers studying the general ocean circulation. Some exist, or have been demonstrated; others show promise and might be developed in the next five years. Most of these developments are unlikely to be used in the field unless there is a broad base of strong support for them.

Examples of such developments with at least a potential for generating useful observations on basin scales to near-global scales are as follows:

- The SEASAT scatterometer. Realistic quantitative estimates of the wind stress and its spectrum are possible from measurements by such an instrument. The problem of directional ambiguity can be overcome with more recent designs.
- The SEASAT altimeter. With the ability to measure absolute sea surface slopes within a few centimeters over spatial scales of 25 km to 10,000 km, the absolute topography and variability of the sea surface of the ocean could be determined globally for periods of days to years.
- Long-range floats. Both neutrally buoyant subsurface floats and surface drifters can now be tracked over long distances. Ocean circulation could be determined at least on ocean basin scales, although the number of floats/drifters necessary remains to be clarified.
- Surface measurements. In-situ measurements of surface air pressure and other quantities yielding air-sea fluxes can be made globally from surface drifting buoys. These could be deployed with spacing compatible with atmospheric synoptic scales to provide the surface boundary condition for the ocean.

• <u>Tracer measurements</u>. Chemical techniques for tracer sampling from small water samples on a routine basis are available for steady and transient tracers, including nutrients, tritium, helium-3, and halocarbons (Freons). For the first time one could obtain from ships global baseline property distributions with mesoscale resolution along sections.

• <u>Acoustic tomography</u>. Existing technology and designs could provide large area integrals (means) of heat content, heat flux, and potential vorticity, as functions of depth over entire ocean basins (across hundreds or thousands of kilometers).

• <u>Eddy-resolving numerical models</u>. These are increasingly realistic over domains approaching those of ocean basins. With anticipated developments in computer technology and numerical methods, such models could be adapted to operate with basin- or global-scale data sets in assimilation modes analogous to those used in meteorology--if the appropriate data existed.

These developments have appeared at the same time that there is growing international interest in the role of the ocean in climate, and climate changes. Thus, international climate research programs, specifically the World Climate Program, could provide the framework for actually carrying out observations of the global ocean. But the fundamental problem is not that of climate per se; it is the problem of determining and understanding the general circulation of the ocean, which must be known to understand the role of the ocean in climate. It is possible that the new developments could be used for observation of ocean circulation, for studies of climate and for other purposes. For these reasons, the Ocean Climate Research Committee of the Board on Ocean Science and Policy appointed a Steering Committee to organize the workshop entitled "Global Observations and Understanding of the General Circulation of the Oceans."

Fifty-eight individuals were invited to participate in the workshop representing physical, chemical, and biological oceanography. The letter of invitation and the workshop agenda are included in Appendix I, and the list of participants is given in Appendix II. Before the workshop, some participants prepared papers on the general concepts of global observations and experiments as well as on the technical aspects of global observations. The compiled background papers were printed and bound prior to the workshop, where they served as a basis for discussion. Because of the importance of these papers to the deliberations, they are included in Appendix III.

In organizing the workshop, the Steering Committee attempted to assure that a wide spectrum of views from the entire oceanographic community would be represented. The result was often heated debate on issues ranging from the concept of global oceanography to differences of opinion about the technical capabilities of particular observational methods. Nonetheless, by the end of the workshop, a consensus did develop that a World Ocean Circulation Experiment (WOCE) appeared feasible, worthwhile, and timely. The workshop participants made no attempt to design such an experiment, but did agree that such a program should have the following overall goal: <u>To understand the general</u>

<u>circulation of the global ocean well enough to be able to predict ocean response and feedback to long-term changes in the atmosphere.</u>

Several other major conclusions were also generally agreed upon: that several of the available and developing technologies made such a goal realistically achievable for the first time and thus that the time is right for planning and executing such a global experiment; that the resources required would demand an unprecedented degree of consensus and cooperation in the world oceanographic community; that the immense effort required to mount such an experiment would be worthwhile; and that such a program would likely require a decadal time scale, with a more intensive central field phase lasting perhaps 5 years beginning about 1990.

To move beyond these generalizations the workshop participants recommended that plans for WOCE should proceed. They are hopeful that the workshop conclusion and support for a U.S. component of the WOCE will be endorsed by the oceanographic community and that the U.S. funding agencies will ask the National Research Council to appoint a committee for the World Ocean Circulation Experiment. The committee would provide regular guidance on planning for a U.S. WOCE component, work with appropriate panels of the Committee on Climatic Changes and the Ocean (CCCO) and the Joint Scientific Committee (JSC) in order to ensure adequate liaison with international planning for the WOCE, and advise the appropriate U.S. government agencies on the implementation of a U.S. WOCE component.

These proceedings reflect views expressed by workshop participants, and the overall goal and specific objectives as stated in Chapter 2 represent a consensus reached by the participants. The suggested next steps in the development of a U.S. component of WOCE are outlined in Chapter 3. The reports of three working groups, which were organized during the workshop to consider specific technical issues, are included in Chapter 4.

CHAPTER 2

STATEMENT OF GOALS AND OBJECTIVES

The workshop participants agreed on this overall goal of the contribution to a World Ocean Circulation Experiment (WOCE):

Overall Goal: To understand the general circulation of the global ocean well enough to be able to predict ocean response and feedback to long-term changes in the atmosphere.

In order to meet that goal, a number of specific objectives were identified by the workshop. These specific objectives are as follows:

1. To complete a basic description of the present physical state of the ocean.
2. To improve the description of the atmospheric boundary conditions on the global ocean and to establish their uncertainties.
3. To describe the upper boundary layer of the ocean adequately for quantitative estimates of water mass transformation.
4. To determine the role of interbasin exchanges in the global ocean circulation.
5. To determine the role of ocean heat transport and storage in the heat budget of the earth.
6. To determine seasonal and interannual oceanic variability on a global scale and to estimate its consequences.

The workshop participants recognized that the goal and objectives of the U.S. component of WOCE are broad. This is intentional. We may not fully attain all of the objectives. However, they constitute a focus and give direction to the program.

In the overall goal is included the ability "to predict ocean response and feedback to long-term climate changes in the atmosphere" in order to have a criterion by which we can judge our progress toward this very difficult task of understanding the global ocean circulation. Through the comparison of global and regional data sets with the results of developing numerical and analytical predictive models, we will attempt to measure our progress.

On the way to attaining the overall goal, a number of valuable intermediate objectives would be achieved. Some of these are identified specifically. The first specific objectives (1 through 3) listed deal

with basic descriptions. These are followed by objectives concerning fundamental processes or phenomena to be understood for the construction or predictive models.

OBJECTIVE 1: To complete a basic description of the present physical state of the ocean.

The basic description of the present physical state of the ocean is meant to include first global distributions of density, passive tracers, and low-frequency horizontal velocity. This description is basic to understanding the circulation of the ocean as a single entity. It may also include estimates of potential vorticity, vertical velocity, and other derived quantities over subregions of the ocean. Program elements essential to this objective are as follows:

- Long-term global observations of sea surface topography by satellites, supplemented by tide gauge measurements;
- Completion of the global sampling of ocean tracers (now under way as the Transient Tracers in the Ocean program);
- Global sampling of the temperature and salinity fields;
- Global sampling of the ocean velocity field by Lagrangian Drifting Buoys; and
- Global sampling of the surface forcing fields by a combination of satellite and surface measurements.

It may be necessary to resurvey the global density field instead of simply filling in present data gaps and repeating density measurements in selected areas. Other requirements may include widely spaced direct measurements of low-frequency currents on a global basis and intensive direct current measurements in selected energetic regions.

OBJECTIVE 2. To improve the description of the atmospheric boundary conditions on the global ocean and to establish their uncertainties.

The second specific objective complements the first objective by providing the boundary conditions needed for the application of models predicting atmospheric forcing effects on the ocean. Moreover, this information provides boundary conditions necessary for the use of observed oceanic distributions of tracers in describing the time-averaged ocean circulation as well as internal mixing.

Specific requirements in support of the second objective include measurement of surface wind stress by scatterometer from spacecraft and other satellite measurements required for global surface heat flux. Global estimates, derived from satellite, surface drifter, and ship data, of radiation, latent heat, precipitation, and accurate sea surface temperature, will be essential to determining surface heat and water fluxes. Efforts are needed to learn how to make these measurements. Surface gas exchange rates and variability of sea ice cover are examples of other requirements.

OBJECTIVE 3. To describe the upper boundary layer of the ocean adequately for quantitative estimates of water mass transformation.

In order to make use of improved descriptions of atmospheric boundary conditions (objective 2), we need to understand the effects of such forcing on the surface boundary layer of the ocean (objective 3). Our description of the surface layer must take into account its variability, including the development of the mixed layer and seasonal thermocline. Special attention is needed in areas where there is deep convection.

OBJECTIVE 4. To determine the role of interbasin exchanges in the global ocean circulation.

In addition to atmospheric forcing, it is the interbasin exchanges that provide the connections between the various ocean basins, leading to a truly global ocean circulation. Monitoring of the transport and property fluxes will be needed through Drake Passage, through the Strait of Gibraltar, and between Greenland and Scotland. Also needed is monitoring of the flow of deep water from the Antarctic and North Atlantic into the major basins, which occurs principally as intense, deep western boundary currents. Deep western boundary current measurements are probably the only effective means of determining the magnitude of deep water circulation. Finally, the monitoring of upper level western boundary currents is needed to constrain the upper water budgets in each basin for model calculations. Both deep and upper level boundary current monitoring is needed for direct heat transport estimates.

OBJECTIVE 5. To determine the role of ocean heat transport and storage in the heat budget of the earth.

It is known now that the ocean heat transport across some latitudes is comparable to the heat transported by the atmosphere. However, with the exception of the zonal transport across 24°N in the Atlantic and a few other locations, we have only rough, and sometimes conflicting, indirect estimates of ocean heat flux. In order to understand the global heat balance, the heat transport and storage in the oceans must be known. Many of the specific requirements needed to determine these are needed also in the support of specifics (objectives 1 through 4). In addition, estimates of intermediate and deep water formation rates are required. The flux of liquid water across the sea surface and the oceanic transport of water are important to this goal.

OBJECTIVE 6. To determine seasonal and interannual oceanic variability on a global scale and to estimate its consequences.

We know that there are strong seasonal cycles in the ocean, but we have not yet had the data base to examine their characteristics on a global basis. We also know that certain signals, such as the Southern Oscillation index (a measure of interbasin sea level pressure differ-

ence), are reflected in very large (at least basin-wide) variability of the oceans. However, we do not know the extent and magnitude of such variability, which is essential knowledge for determining the casual mechanisms. Extensive additional sampling probably will not be required to address objective 6, but much can be learned from the requirements for meeting the other objectives.

Scale interactions are not stated as a specific objective, but can be addressed through the opportunity of WOCE. In particular, interactions between eddies and the mean flow, and interactions between western boundary currents and the interior flow of oceanic gyres are essential to the global problem.

Further elaboration of the scientific needs behind the program objectives and of the specific requirements needed to approach them are given in the reports of the three workshop study groups, which are presented in Chapter 4.

CHAPTER 3

NEXT STEPS TOWARD A U.S. WOCE

The main purpose of the workshop was to examine the feasibility and desirability of designing and carrying out a global ocean circulation experiment. The participants agreed that the experiment appeared feasible, and the next necessary steps were considered.

Here are outlined the next steps necessary for the development of a U.S. component of the World Ocean Circulation Experiment as identified by the workshop Steering Committee.

1. Distribute workshop report and solicit comments. This report is being widely circulated within the U.S. scientific community to provide potentially interested participants in a global experiment with the background and conclusions of the workshop. It is to be stressed that this report is not a scientific plan. It consists of inputs to and ideas of workshop participants. Recipients are being asked to provide their thoughts on the questions addressed to workshop participants and to comment on the workshop results. The responses will be used as part of the planning process.

2. Form a U.S. WOCE Committee. A U.S. WOCE Committee should be established within the National Research Council to provide oversight for further planning of a U.S. component of WOCE, to ensure adequate liaison with international planning for the WOCE, and to advise appropriate U.S. funding agencies on the implementation of the U.S. component of WOCE.

3. Establish WOCE working groups. An International Steering Committee for WOCE has been established under the auspices of CCCO/JSC. That committee has established several working groups for the consideration of aspects critical to further WOCE planning, with emphasis upon climatic goals. It is planned to establish U.S. working groups to similarly assist in the formulation of the U.S. contribution. It is suggested that the working groups focus on the following topics:

   i. Ocean surface layer and atmospheric exchanges
   ii. Ocean velocity
   iii. Global temperature and salinity distributions
   iv. Tracer distributions
   v. Oceanic heat flux

vi. Interbasin exchanges
vii. Modeling and data synthesis
viii. Technology development

The first four topics are being addressed also by working groups of the International Steering Committee for the WOCE. (Topic i is meant to include consideration of surface layer processes. Topic ii should include consideration of (1) mean velocities on large scales, (2) time-varying velocity on large to intermediate scales, (3) mesoscale statistics, and (4) scales that might be obtained from topographic measurements by satellites. Topic iv must include considerations of interior mixing.)

Heat flux and interbasin exchange studies should be considered, and initial plans for such subprograms should be formulated.

4. Draft a summary scientific plan for U.S. WOCE. In parallel with international and U.S. working group activities, the U.S. WOCE Committee should begin preparation of a scientific plan for the U.S. component. Because of the many technical and scientific uncertainties bearing on program design, this plan must be preliminary in nature and will continue to evolve over a period of several years. The plan must also take into account any comments, criticisms, or suggestions arising from the workshop report.

CHAPTER 4

REPORTS OF THE WORKING GROUPS

INTRODUCTION

The following three reports, written during the workshop, present the deliberations of the working groups in which the workshop participants considered the following questions:

1. What would constitute an adequate understanding of the general circulation of the ocean?
2. Can we make a major advance in our understanding of the ocean circulation with a global study in the next two decades?
3. What are the critical problems (primary and secondary) that must be solved in order to make this advance (consider your definition of global)?
4. What are the specific physical processes that need to be studied in order to understand the critical problems? Can they be listed in priority order?
5. What data ought to be collected, and what is it realistic to expect in the future? What data fields are required for a proper description? What analysis field? What quantitative approaches need to be carried out to simulate various observational nets (e.g., particular float distribution, particular tracer measurements, various mix of satellite measurements)?
6. What specific model development is needed now, and what developments are to be expected in the future? Consider OGCMs, EGCMs, and data assimilation models.
7. What are the needs for measurement techniques? Do they now exist, or can they be developed in time?
8. What is the strategy for proceeding? Formulate an overall timetable. What studies must be simultaneous? What needs to be done now for planning?

GROUP 1: WATER MASSES AND THEIR EXCHANGE

Members: William J. Jenkins, Co-Chairman
Joseph L. Reid, Co-Chairman

Bert Bolin, Neil L. Brown, Kirk Bryan, Curtis A. Collins, James Crease, Robert Etkins, Manuel E. Fiadeiro, Joseph Pedlosky, Peter B. Rhines, Claes G.H. Rooth, Jorge L. Sarmiento, John Steele, Henry Stommel, Bruce Warren, and Ray F. Weiss.

Motivation

The ocean plays a crucial role in the long-term maintenance of global climate. In addition to transporting and storing heat, it exerts control over atmospheric concentrations of the climate modifier CO_2. The transport of CO_2 and other environmentally important substances to and from the deep ocean occurs on time scales ranging from years to millennia, that is, on time scales most relevant to mankind's interests. Further, the principal means whereby the surface ocean (and hence the atmosphere) communicates with the deep ocean is via the process of water mass conversion, that is, the thermodynamic modification of surface water properties and subsequent vertical motions. Inasmuch as water mass conversion or formation is subject to climatic forcing and its perturbations, there exists a feedback between climate and CO_2 transport. The sense of this feedback may well be positive; that is, the climatic effects resulting from an increase in atmospheric CO_2 concentrations may reduce the ocean's ability to remove the CO_2.

Clearly, then, it is important to understand the processes of water mass formation and motions from the viewpoint of understanding long-term global climate. The difficulty arises in the fact that these processes are sporadic in time and heterogeneous in space. Few direct observations of water mass conversion actually exist, and its occurrence is usually inferred from observations made after the fact. The most powerful tool we have in this regard is the observation of tracers, and in particular transient tracers.

Tracers have some very useful attributes. Dependent on boundary conditions, time history (if transient), and in-situ behavior (e.g., half-life), a tracer will integrate over space and time scales and provide a representative average of the net effects of a given process. Certain tracers, because of their nonuniform boundary conditions, provide visibility to some important processes or features (e.g., tritium in the deep western boundary current in the North Atlantic). Finally, the broad variety of boundary conditions, characteristics, and time histories embodied in the tracers measurable today should provide a means of discerning between various process models that might otherwise be difficult to separate. This last feature is particularly important for model validation and parameterization. Programs such as GEOSECS and TTO (Transient Tracers in the Ocean) have provided high-quality tracer data that have and will be used in models or ocean circulation and transport.

It should be noted, however, that the usefulness of a given tracer is directly dependent upon the extent to which its boundary conditions, time-testing, and in-situ behavior are known and understood. Consequently, an effective observational program may include some activities aimed more at understanding the nature of some of these tracers. To some extent the behavioral characteristics of a tracer may be "backed-out" from observable distributions, but some boundary or process experiments may need to be done. Examples of such experiments include sediment trap and benthic chamber deployments. Consideration of the logistics of such experiments suggests that these experiments may best be pursued in parallel with WOCE activities.

A specific goal of WOCE is to characterize on space scales from basin to global the distribution and transport of heat, mass, and tracers. The ultimate product of this activity would be a successful model(s) of the ocean. A successful model not only should satisfactorily explain all observables, but also should have prognostic capacity: the degree of success in the latter is gauged by the magnitude of the perturbation in forcing that such a model can realistically accommodate. The strength of the model in this regard is directly related to the degree to which it is constrained and validated by observations.

In practical terms, we should not only go about creating a globally "synoptic" data set that will serve as diagnostics for ocean transport models, but also investigate via well-thought-out process-oriented studies the parameterization of those models. A further approach would be to quantify temporal variations and their response to existing forcing changes.

Specific Recommendations

Tracers provide extremely useful information about oceanic transport processes on the climatically important time scales (years to centuries). They provide qualitative and quantitative integral constraints on the crucial questions of water mass creation, transport, and dissipation. The opportunities presented by the WOCE concept, namely, the close coupling of tracer measurements with an intense physical observation program and satellite observation, should be exploited thoroughly. The working group urges the following:

(a) that there be a carefully designed tracer and hydrographic measurement program that is an integral part of the WOCE program. The relationship between WOCE and the TTO program should be carefully examined in this context;

(b) that the scope of the WOCE program include not only large-scale quasisynoptic surveys, but also time series measurements in key places at the appropriate frequencies, and regionally intensive studies aimed at improving our knowledge of specific processes and boundary conditions (both chemical and physical), an example being high-latitude water mass formation regions;

(c) that some attempt be made to coordinate the tracer and hydrographic measurement program with such parallel activities as float and drifter releases and current meter deployments;

(d) that some consideration be given to better establishing the behavior (boundary conditions, time history, and in situ behavior) of some tracers that may be of use to the overall WOCE effort;

(e) that some attempt be made to optimize sampling strategies on the basis of existing models, and to foster model development on the basis of existing data; and

(f) that a panel or working group be assembled immediately with the charge of following through on the above recommendations.

Tools

Tracers represent the primary tools for studying water mass formation, motion, and mixing, but clearly we need to start with a well-defined and carefully monitored foundation of high-quality hydrographic measurements. Measurement of the field of mass should be a fundamental component of this program.

The more useful tracers from the viewpoint of examining ocean transport processes are primarily those tracers that have no (or very weak) in-situ biological or chemical activity. These include the isotopes produced by bomb testing: tritium (^{3}H)- (^{3}He, ^{90}Sr); industrial byproducts: ^{85}Kr, halomethanes (Freons); and certain natural isotopes: ^{39}Ar and ^{228}Ra. Other tracers of biogeochemical interest include the nutrients (nitrate, phosphate, and silicate), oxygen, CO_2, and ^{14}C. This latter group also provides semiquantitative constraints on transport and mixing and, where their source/sink terms can be determined with some precision, may prove valuable diagnostics of ocean models. Other tracers may eventually be added to this arsenal, depending on agreements in measurement techniques and our understanding of their behavior.

First-order observations of tracer distributions and their evolution afforded by the GEOSECS and TTO programs provide important guidance and constraints in the formulation of models and theories of ocean transport. Beyond this, the interrelationships among the tracers are useful and potentially definitive tests of these models. With this in mind, a well-planned regional study should not only emphasize those tracers with the most appropriate boundary conditions, but also include other tracers (at a more modest level) that provide ancillary or sometimes even redundant information.

The tracers mentioned above can be divided into two groups on the basis of logistics: those tracers that require relatively small volumes (≤ 10 liters) and those that entail the acquisition of large (≥ 50 liters) volumes of water.

The former (small volume) group may be further divided into those tracers that are measurable on the ship (e.g., O_2 nutrients, Freons, and CO_2) and those that are measured on shore (e.g., ^{3}H and ^{3}He).

The large-volume group includes ^{90}Sr, ^{14}C*, ^{85}Kr, ^{39}Ar, and ^{228}Ra. Whereas it is tempting, and indeed more expedient, to focus on the small-volume tracer group, the information offered by even a coarse, relatively infrequent sampling of the large-volume sample provides a powerful incentive to support some measurement activity in this area.

Finally, it should be recognized that although the techniques differ in practice, the combination of a tracer measurement program with limited float and drifter releases on current meter deployment may prove a useful approach in certain oceanic regions. For example, a detailed characterization of tracer distributions in a boundary current during a current meter record may allow a direct estimate of fluxes. In addition, the use of remote sensing to obtain a large-scale average of the sea surface state during periods of water mass formation may complement tracer measurements in studying air-sea exchange on a regional basis.

Areas of Interest

The first goal of WOCE will be to obtain a global map of the state of the ocean at one time. The fact that such a picture cannot be truly synoptic is a concern, but a sensible sampling strategy extending over decadal time scales can fulfill this goal when coupled with the satellite data and some repeat observations.

Clearly, the coverage of the ocean will not be uniform: certain areas of the ocean will require more intensive sampling (e.g., the Gulf Stream) and perhaps even repeated measurements to examine secular and seasonal trends. Although special thought should be given to ensuring that gaps in the existing data base be filled in, resampling areas that are already well characterized by modern hydrographic standards may be especially useful in assessing the representativeness of the WOCE surveys.

Areas that will require this more intensive sampling include those places where transport is constrained either dynamically or topographically to narrow flows. These include boundary currents (e.g., the Gulf Stream and Kuroshio), major overflows (e.g., Denmark Straits and Strait of Gibraltar), and channels (e.g., Drake Passage and Vema Channel). Regions where major water masses are formed may be the sites of repeated sampling or process studies. These include thermocline outcrop regions (e.g., Northeast Atlantic), intermediate water mass "breeding grounds" (e.g., Labrador Sea and Southeast Pacific), and the polar seas. Finally, upwelling regions (e.g., the circumpolar and equatorial regions) may need special attention.

It is impossible within the scope of this document to fully formulate the sampling strategy. More work is clearly needed as planning progresses to develop a viable program.

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*It should be noted that with improvements in acceleration techniques, it may become possible to measure ^{14}C on a small-volume sample with sufficient accuracies for ocean tracer purposes.

Experimental Approach

There are four observational styles, each being not necessarily exclusive of the other, that will prove valuable to the WOCE program. The philosophy of large-scale surveying of tracer and hydrographic properties is the most important of the four. Time series sampling, however, may provide important correlative information on higher frequencies, and some process-oriented studies may give essential guidance in the interpretation of the "synoptic" data. The relative mixture of these approaches needs to be carefully thought out, and should be considered an important component in planning.

1. Large-scale quasisynoptic sampling. Measurement of the large-scale features of oceanic property distributions (both passive tracers and dynamically active properties) is the cornerstone of the WOCE concept. There is a need to fill in some "data gaps," as well as to look for secular changes in the more conventional observables: temperatures, salinity, density, dissolved oxygen, and some nutrient salts. Further, the delineation of the distribution of more "exotic" tracers, particularly transient (man-made) tracers with moderately well understood boundary conditions, will prove an important step forward in our knowledge. Such tracers will prove critical diagnostics of ocean models that attempt to describe water mass creation, transformation, and transport.

One component of such a program should include a TTO-style program of relatively coarse-grid sampling of both small-volume and large-volume tracers. This should be coordinated with a more detailed survey of hydrographic/nutrient distributions on a scale suitable for dynamical studies. This second program should include a subsampling of small-volume tracers to fill in the coarser grid tracer patterns.

2. Serial sampling. Observation of the variability and time evolution of certain properties will be important in two ways. First, it will aid in the interpretation of the not-perfectly-synoptic large-scale data set. Second, such variations are in part a result of temporal variations in climate forcing, and hence important observables in themselves.

The existence of spatially homogeneous "provinces" in hydrographic and tracer distributions makes it possible to characterize regional trends and variations with a relatively modest monitoring effort. A carefully laid out sparse monitoring network would not only be scientifically effective, but would also be amenable to international collaborative effort. The frequency of sampling depends both on the region and the properties measured and on logistic constraints. One would picture a minimal sampling frequency of the order of one per year, and probably shorter in some places where seasonal effects need to be studied.

A "pre-WOCE phase" should include the start-up of such sampling programs as deemed appropriate, and they should be continued through a "post-WOCE" phase. This may provide important constraints on the representativeness of the WOCE data with only a modest expenditure of effort.

3. Regional (process-oriented) studies. Certain critical areas will require more highly resolved and intensive measurement programs. Some carefully designed experiments may do much to improve our knowledge of boundary conditions or in-situ behavior of some tracers. This may, in turn, make the large-scale distribution more useful. Care must be taken, however, to decide how much of this is appropriate within the WOCE program, and how much is better pursued in parallel programs.

4. Purposeful injections. The deliberate injection of significant quantities of environmentally safe substances into the ocean so as to observe their subsequent dispersal on large space and time scales may be a powerful approach to studying ocean transport processes. A number of different substances may be introduced at a number of places and times, and the boundary conditions (i.e., behavior at the ocean surface or sides, etc.) may be used in varied ways to illuminate the different processes occurring. Clearly, substantial engineering and technical problems must be overcome, and the scientific design of the experiment (and the interpretation) must be careful, but there is promise in the technique.

Modeling and Interpretation

Theoretical and numerical model development is important within the framework of WOCE for two reasons. First, the existence of a "working hypothesis" will prove valuable in formulating and optimizing a sampling strategy. Second, the task of managing and assimilating the data that will result from such a program may be aided by working models.

Both eddy-resolving and non-eddy-resolving numerical models have shown progress in the past decade due to improvements in computing power and theory, and further improvements are expected. Development of primitive equation "thermodynamic" models that couple to atmospheric forcing in a realistic way is needed, and is under way in some groups. Regardless of the approaches used, such models need to be formulated in a way that makes them amenable to validation by actual observations.

Theoretical models have been developed to describe the morphology of wind gyres and the injection of tracers into them from the atmosphere. The equivalent theoretical and numerical studies of the mid-depth and abyssal circulation are much more difficult. We are coming to an understanding of the circulation riding upon a known basic stratification (a 10-year problem in terms of response times), but the production of the basic stratification (a 1000-year problem) is the challenge that parallels the observational parts of WOCE.

Some rudimentary models exist, and some effort should be expended in exploring their utility in experiment design and validation. Techniques of objective mapping and resultant error fields may be useful in that regard. Beyond this, work is already progressing within independent groups. The strategy for WOCE might be to support and stimulate communications between the modeling groups and the investigators involved in the field work.

Summary

Water mass conversion and motion is an important process in the ocean-atmosphere-climate system. The best tool for studying this process is the observation of transient and steady-state tracer distribution. We recommend the planning of a large scale tracer and hydrographic sampling program coupled with some time series and process-oriented studies.

GROUP 2: ATMOSPHERE-OCEAN EXCHANGE

Members: Pearn P. Niiler, Co-Chairman
William George Large, Co-Chairman

Joost A. Businger, Anthony Calio, D.E. Harrison, Robert H. Heinmiller, Jr., A.D. Kirwan, Ryuji Kimura, John Morrison, William C. Patzert, James F. Price, Roger Revelle, Robert H. Stewart, R.W. Stewart, and John Woods.

Specific Goals

1. To improve the description of the atmospheric boundary conditions on the global ocean and to establish the uncertainties in each of them.
2. To obtain a global description of the upper boundary layer of the ocean (including the seasonal thermocline and focusing on areas of deep and intermediate convection) adequate for quantitative estimates of water mass transformation therein.

Issues and Recommendations

From the Background Papers

Most of the pertinent issues concerning atmosphere-ocean exchanges from a WOCE viewpoint are contained in the prepared background papers of Large, O'Brien, and Woods (see Appendix III). The most relevant points are listed below for reference, but the papers themselves should be consulted for details and context.

- For global observations, satellites will be necessary but not sufficient, as surface observations are still required for some parameters (e.g., air temperature), for calibration data, and for continuous time series.
- The SEASAT experience demonstrated the capability of a scatterometer to measure winds and that geographical averaging could substantially reduce the random error. The frequency-wave number spectrum of the surface wind stress is needed for a quantitative assessment.

• The direction ambiguity problem of SEASAT has been much reduced for future scatterometers, but there is still the issue of whether it is wind speed or stress that is being measured.

• All satellite measurements will require corroborating surface data in order to reach their potential accuracies and hence usefulness. It is essential that plans be made to perform the necessary experiments in a variety of climates.

• If scatterometer measurements of wind stress and velocity prove good enough to be considered as a standard measurement, then scatterometers will provide an opportunity to learn how to estimate stress to WOCE requirements from other data (e.g., underwater acoustic noise, altimeter data, sea level pressure) as well as from indirect refined techniques such as atmospheric model outputs and those using heavily filtered data. It would then be possible to continue stress measurements when there are no scatterometers, though only satellite cloud motion vectors in the tropics and a pressure network and analysis elsewhere provide directional information. Such a network would also affect our ability to extend stress estimates into the past. It is recommended that preliminary calculations be initiated using existing data and that plans be made to exploit the next scatterometers.

• The uncertainty in global bulk aerodynamic estimates of the turbulent exchanges is now due more to bulk variable measurement errors (e.g., in wind, SST, and humidity), than to the transfer coefficients C_D, C_E, and C_H. Tuning the magnitudes of these coefficients in order to compensate for systematic measurement errors is not a satisfactory procedure.

• The net solar radiation (now obtainable from satellites) and the latent heat flux are the two most important components of the surface heat flux, but neither the longwave radiation nor the sensible heat flux can be neglected everywhere all the time.

• Present methods of estimating globally the annual net surface heat flux have a very large uncertainty. Therefore new technology, refined and innovative techniques, and new ideas and data sources (atmospheric model output?) need to be explored. What will be the impact on WOCE of a great improvement? of little improvement?

• Our second specific goal requires knowledge of the seasonal variation in the heat flux components, the solar heating profile, and precipitation. All of these influence the seasonal cycle of mixed layer depth, temperature, and salinity, and hence water mass conversion. This knowledge will also be needed to test coupled ocean-atmosphere models of the planetary climate system.

• For WOCE, the upper boundary layer is an important buffer between the atmosphere and the deep oceanic circulation. The importance of, and techniques for estimating, the depth and extent of late winter convective overturning needs to be evaluated.

From the Discussions

Issues raised and recommendations made by Group 2. These items for consideration are in addition to those in the background papers and are not in order of priority.

• Observational strategy. The design of the global observation network to provide "adequate" descriptions of the important ocean and atmospheric boundary layer parameters must be well-matched to the overall goals of WOCE. Thus, careful consideration of the spatial and temporal scales of the most important processes must be defined. Examples are the annual signals in mid- and high-latitudes, the role of storms in both hemispheric winters, and the lower-frequency (inter-annual) variability, not only in the tropical but also over the southern ocean. Our knowledge of the southern ocean is primitive in relation to that of the northern, but its description is crucial to several of the WOCE goals.

Feasibility/error/design analyses are needed that reconsider the mix of satellite and in-situ boundary layer observations needed to adequately define and calculate the heat budget of the global oceans. Can we estimate heat budgets adequately? If not, which parameters are important and crucial: SST, wind speed, wind stress, total or boundary layer water vapor, short-wave radiation reaching the sea surface, etc.? Ships-of-opportunity can provide "lines," drifting "flux" buoys can provide moving points, and so on. Serious uncertainty exists as to the accuracy of satellite products, such as SST, though they certainly have the most promise. The application of satellite data to the description of the marine boundary layers needs to be studied and limits, errors, and problems be established. The options open to WOCE range from quantitative to qualitative descriptions and from global coverage to a focus on regional "adequately" sampled experiments.

• Wind stress. The most important exchange is that of momentum (surface stress), and it does seem that it will be possible to make global observations using primarily a scatterometer, but supplemented by other satellite data (e.g., altimeter). A great deal could be done to improve and assess scatterometer stress accuracy, the most significant improvement being to relate backscatter directly to stress rather than to wind speed. However, before the required effort can be determined and initiated, the accuracies needed for WOCE have to be specified. It is recommended that this be done soon.

• Heat exchange. At present, global satellite measurements of the net solar heating are possible. There is an ongoing attempt to estimate the latent heat flux, which has had some encouraging success, but a serious evaluation of the method is needed. Development of satellite longwave radiation techniques is about to begin, but as yet there have been no attempts to estimate the sensible heat flux from satellites.

The satellite components of WOCE need to be expanded from the altimeter and scatterometer to include the other satellite measurements required for global surface heat flux if these measures and surface circulation are to be given a high WOCE priority. Global satellite

radiation, latent heat, precipitation and SST estimates will be required for determining the surface heat and water fluxes. Efforts need to be initiated with the goal of learning how to make these measurements, and perhaps how to either infer or measure directly the total heat flux.

Related to boundary layer conditions is the heat content of the upper oceans. Over certain regions of the ocean and times of year, it can be monitored with ships-of-opportunity expendable bathy thermograph (XBT) programs and in a primitive manner with altimeter data. Over certain large areas and certain seasons, its changes are closely related to the net surface heat flux.

The accurate calculation or the solar heating profile in the ocean is required for estimation of water mass conversion below the mixed layer in the tropics. Sensitivity tests have shown that it is necessary to take account of the turbidity (i.e., phytoplankton concentration) of the upper ocean. This might be done globally by suitable processing of data from a satellite ocean color imager. A program of in-situ measurements will be needed for calibration.

• Liquid water flux. The net flux of liquid water at the surface acts as a salinity condition, and this is an important boundary condition for climatic water mass formation problems. The flux is determined by precipitation, ice processes and evaporation and is poorly known at present. Neglecting ice processes, the major problem is the difficulty of obtaining quantitative precipitation information. Precipitation is extremely intermittent in space and time and thus hard to sample. All means of improving precipitation measurements should be considered, especially those using satellite radiometers, because they are showing some promise (SEASAT, Scanning Multichannel Microwave Radiometer (SMMR)).

• Trace gas exchange. A most important climate factor is the influence of anthropogenic production of CO_2 and other trace gases in the atmosphere. There is evidence to suggest that the $_pCO_2$ level in the oceanic mixed layer importantly affects atmospheric $_pCO_2$ and other gases may behave similarly. Although not central to WOCE, its observational network should be equipped to make mixed layer $_pCO_2$ measurements, and the possibility of improving our knowledge of gas exchange rates, and of utilizing a satellite ocean color instrument, should be examined.

• Critical regions. A close look needs to be taken to determine critical study areas where special efforts are needed to describe the surface forcing functions and boundary conditions. These include (1) regions where the boundary exchanges are particularly strong, such as the large evaporation from the Norwegian Sea, where sea surface temperature cannot be determined by conventional satellite means, (2) regions where few or no data are currently available for evaluating satellite products, such as the South Pacific Ocean, and (3) regions where the depth of summer thermocline is much shallower than the depth of winter penetrative convection.

• Ice coverage. It is recognized by this group that the effects of seasonal and interannual variability of sea ice coverage on the atmospheric boundary conditions on the global ocean need to be studied

(see paper by Woods in Appendix III). This is a very important component of the exchange between the atmosphere and the ocean, which is now being monitored routinely from satellites. It is necessary to determine whether these data are sufficient, and if not, what else is required.

GROUP 3: VELOCITY FIELD

Members: Harry L. Bryden, Co-Chairman
Walter H. Munk, Co-Chairman

Robert E. Cheney, Russ E. Davis, Michael Ghil, Dale E. Haidvogel, Michel Pierre Lefebvre, James R. Luyten, James G. Marsh, Paola M. Rizzoli, Allan R. Robinson, Robert C. Spindel, George Veronis, Douglas C. Webb, and Stan Wilson.

This report is divided into two sections: the first deals with a distributed system contributing to a reference level velocity on a global scale; and the second, the monitoring of currents in selected critical regions. Both strategies contribute to the basic description of the physical state of the global ocean and constrain model predictions. It is our intention to list only those measurements that would be extremely useful but fall outside what we consider to be the scope of WOCE.

Reference Level Velocity

We list what we consider the vital components of a velocity measuring program. The list begins with those of a truly global distribution and ends with those covering certain more restricted regions.

1. Satellite altimetry yielding surface geostrophic currents. We expect some important improvements in our knowledge of the geoid prior to deployment.
2. Surface drifters with daily satellite fixes to give total surface currents. Here there are problems of slippage from wind drag that need further study.
3. Pop-up buoys with monthly (say) satellite fixes to provide integrated velocities at selected depth(s) down to 1 km.
4. Acoustic Doppler measurements from vessels to provide upper-ocean shear measurements; with precise navigation these measurements give absolute velocity profiles.
5. SOFAR floats with daily (say) acoustic fixes at selected depth(s) between 0.5 and 2.5 km. Moored autonomous receiver moorings at 1000-km intervals are required.
6. Tomographic arrays to provide gyre-scale averages of velocity, vorticity, and temperature. Two such arrays are now envisioned. The averages are Eulerian (techniques 2 through 5 are Lagrangian). Some combinations of (5) and (6) can provide variable array geometries.

7. Moored current meters at critical locations. These are particularly useful for deep currents.

The systems listed here are in various stages of development. None are in a state of readiness; all of them have been tested (or partially tested) at sea. A most intense effort between now and 1989 is required.

Critical Region Monitoring

Monitoring the flow in critical regions can help to define not only the concentrated ocean currents but also the magnitude of the broad, sluggish return flows over large mid-ocean regions of the ocean. We have divided the critical regions into three categories: overflows, western boundary currents, and interbasin exchanges.

In the category of overflows, we include the Denmark Straits-Faroc-Scotland overflow of North Atlantic deep water from its formation region in the Norwegian Sea into the North Atlantic and the Mediterranean outflow through the Strait of Gibraltar into the subtropical North Atlantic. Monitoring these overflows should yield powerful constraints on the deep and intermediate level circulation in the North Atlantic.

Western boundary currents, both upper level and deep, should be monitored for five years during WOCE. We envision this as an international effort. The upper level western boundary current transport and its temperature and salinity characteristics should be defined in each ocean basin: the Gulf Stream flow through Florida Straits, the Kuroshio flow through Tokhara Strait, the East Australian Current off Australia and New Zealand, the Brazil Current off Brazil or Argentina and the Somali Current off east Africa and Madagascar. These measurements will severely constrain the upper water circulation in each ocean basin for model calculations. In addition, we recommend that the large eastward Gulf Stream flow south of New England also be monitored as a further constraint on the North Atlantic circulation. Deep western boundary current measurements are probably the only effective means of determining the magnitude of the deep water circulations. It is especially important that the North Atlantic deep water flow against the continental slope of eastern North America be measured. We also think that small critical regions might be found to measure the northward flow of Antarctic bottom water in the South Atlantic, Pacific, and Indian oceans. This combination of upper water and deep western boundary current measurements is also necessary for good estimates of neat and freshwater transports in each basin.

Finally, for interbasin exchange, we identify the critical region as Drake Passage where the Antarctic Circumpolar Current is most severely constrained. A determination of the transport through Drake Passage and its variability should provide strong constraints on the entire circumpolar circulation. We would also like to have defined a set of measurements to determine the cross-equatorial flow between the North and South Pacific, Atlantic, and Indian oceans, but we are intimidated by the zonal scale of the equatorial zones and a fear that

the velocity correlation scale would be very short. We recommend, however, that some preliminary scale definition measurements be made in some equatorial ocean to determine if a measurement of the cross-equatorial flow is feasible.

Figure 1 shows a map of the primary regions of critical flow monitoring. In each of these regions, we envision a five-year monitoring of the flow. Finally, we see many opportunities for making these measurements with cost-effective monitoring schemes that are currently being developed.

Other Considerations

These measurement techniques, which are central to WOCE, should be available by 1987 at the latest so that an internationally coordinated field experiment can take place beginning in 1989.

We see these velocity measurements as an international program. There are two good reasons for this: first, it is too big a job to be undertaken by a single nation; and second, it offers an excellent opportunity for collaboration between oceanographers around the world.

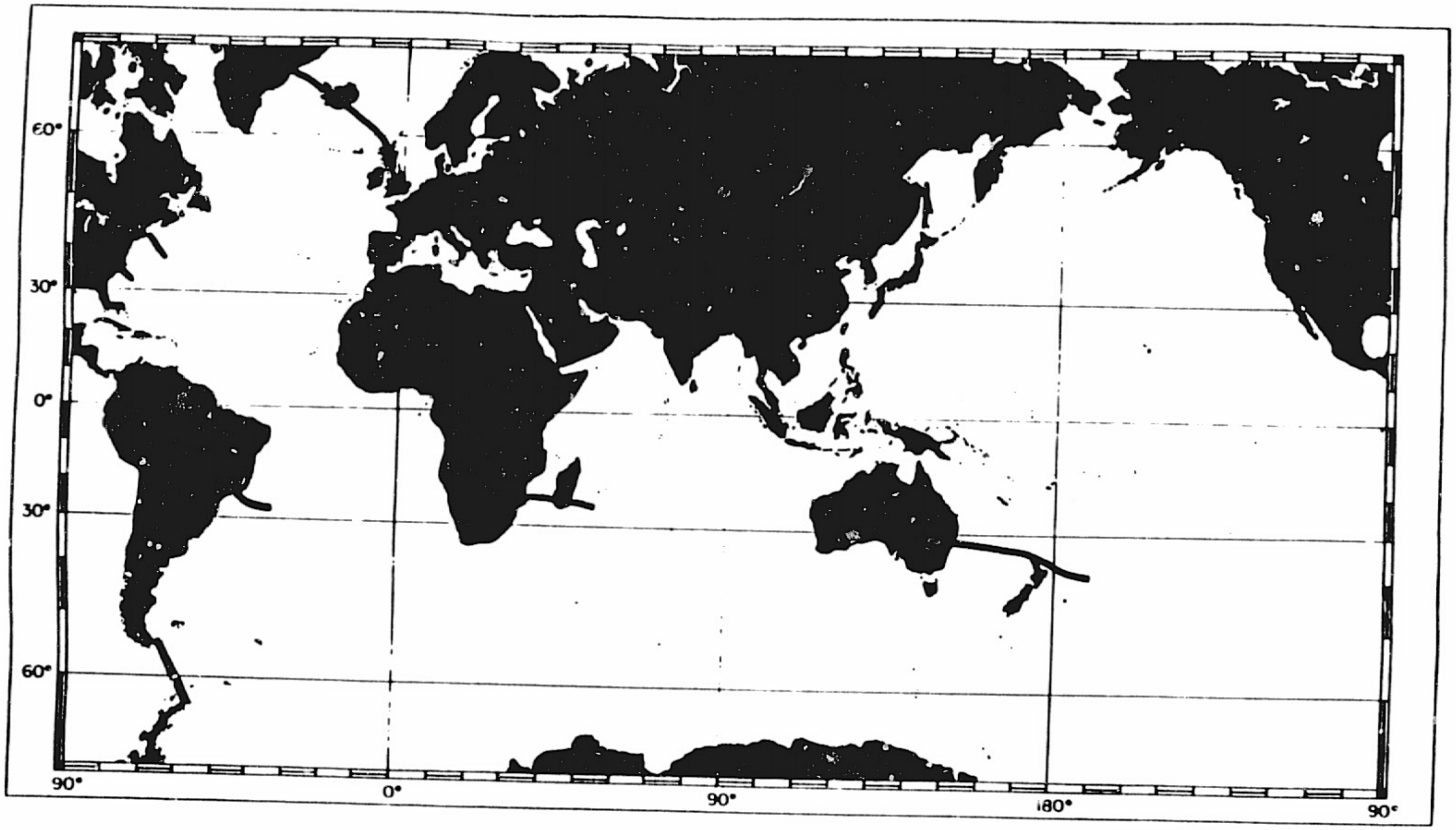

FIGURE 1 The primary regions of critical flow monitoring.

APPENDIX I:

AGENDA AND
LETTER OF
INVITATION

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AGENDA

Workshop on Global Observations and Understanding of the General Circulation of the Oceans

National Academy of Sciences
Woods Hole Study Center

August 8-12, 1983

<u>SUNDAY</u>
7 August

Steering Committee Meets 10:00-14:00 at Academy
16:30 Academy Party

<u>MONDAY</u>
8 August

9:00 a.m.	Introduction and Logistics of Meeting (C. Wunsch, M.H. Katsouros)
	Discussion of Agenda and Purpose of Meeting
	Concepts of World Ocean Circulation Programs (F. Webster, Moderator) (C. Wunsch, J. McWilliams, F. Bretherton)
10:15 a.m.	Coffee
12:15 p.m.	Lunch
1:15 p.m.	Concepts of World Ocean Circulation Programs (Continued) (J. Baker, Moderator) (W. Broecker)
	Discussion
3:15 p.m.	Coffee
3:30 p.m.	Overview of Background Technical Papers 1. Ocean Observations Drifters (surface/subsurface) (R. Davis, P. Niiler)
4:30 p.m.	Adjourn

<u>TUESDAY</u>
9 August

8:30 a.m.	Overview of Background Technical Papers (J. McWilliams, Moderator) 1. Ocean Observations (Continued) Hydrography - W. Nowlin and J. Reid Chemical Tracers - R. Weiss and W. Broecker Sea Level Tide Guages and Altimeters) - C. Wunsch Developing Technologies - R. Heinmiller
10:00 a.m.	Coffee
10:30 a.m.	Overview of Background Technical Papers (Continued) 2. Air-Sea Transfer and the Forcing Functions The Wind Field-Satellite - F. Bretherton and R. Stewart The Wind Field-Conventional - W. Large Thermodynamic Coupling - P. Niiler and W. Large Seasonal Cycle - J. Woods Developing Technologies - R. Heinmiller

12:00 NOON Lunch

1:00 p.m. Overview of Background Technical Papers (Continued)

3. Modeling
   Large Scale Numerical Modeling - D. Haidvogel and K. Bryan
   Assimilation Problems - N. Phillips
   Use of Chemical Tracers - J. Sarmiento

3:00 p.m. Coffee

3:30 p.m.

4. Theoretical Perspective - J. Pedlosky, et al.

5. Critical Regions - C. Wunsch

Discussion

4:00 p.m. Adjourn

6:00 p.m. Clambake at Study Center

WEDNESDAY
10 August

8:30 a.m. (CARRIAGE HOUSE)

Plenary

Overview of Background Technical Papers (Continued)

10:15 a.m. Coffee

10:30 a.m. Working Group Session:

Working Group 1 - Water Masses and Their Exchange
Co-Chairmen: W. Jenkins and J. Reid

Working Group 2 - Atmospheric-Ocean Exchange
Co-Chairmen: P. Niiler and W. Large

Working Group 3 - Velocity Field
Co-Chairmen: W. Munk and H. Bryden

Working Groups Meet to Consider Questions

12:00 NOON Lunch

1:00 p.m. Working Group Session

3:15 p.m. Coffee

3:30 p.m. Plenary

5:00 p.m. Adjourn

<u>THURSDAY</u>
11 August

8:30 a.m.	Plenary
	Report by Co-Chairmen of Working Groups
10:30 a.m.	Coffee
10:45 a.m.	Working Group Session
	Working Group 1 - Carriage House Working Group 2 - Room 202 Working Group 3 - Room 204
12:00 NOON	Lunch
1:00 p.m.	Working Group Session Continues
2:30 p.m.	Coffee
3:00 p.m.	Plenary
	Discussion of WOCE Major Goals and Programmatic Elements
4:00 p.m.	Adjourn
6:00 p.m.	Dinner Party at Wunsch Home

<u>FRIDAY</u>
12 August

8:30 a.m.	Plenary
	Discussion of Working Group Reports
12:00 NOON	Adjournment of Workshop
	Lunch

NATIONAL RESEARCH COUNCIL

COMMISSION ON PHYSICAL SCIENCES, MATHEMATICS, AND RESOURCES

2101 Constitution Avenue Washington, D. C. 20418

BOARD ON OCEAN SCIENCE AND POLICY

OFFICE LOCATION:
JOSEPH HENRY BUILDING
21ST STREET AND
PENNSYLVANIA AVENUE, N.W.
(202) 334-2714

LETTER OF INVITATION

Dear

The Ocean/Climate Research Strategies Committee of the Board on Ocean Science and Policy (BOSP) and in cooperation with the Board on Atmospheric Sciences and Climate of the National Research Council will convene a Workshop on Global Observations and Understanding of the General Circulation of the Oceans, to be held August 8-12, 1983, at the National Academy of Sciences Study Center, Woods Hole, Massachusetts. On behalf of the Boards, I cordially invite you to participate in this workshop. The purposes of the workshop are: (1) to determine whether we have the ability to obtain ocean data on a global scale that could profoundly change our understanding of the circulation; (2) to identify the primary and secondary elements needed to conduct a global ocean circulation experiment; (3) if the ability is achievable, to determine what the U.S. role in such an experiment should be; and, (4) to outline the steps necessary to assure that an appropriate program is conducted.

Before the workshop, a number of experts will be expected to prepare "white papers" addressing technical issues such as the specifics of a satellite wind measurement program, complete with accuracies, coverages, and use of the data. These papers will provide a straw man for workshop participants and will be distributed prior to the workshop. Workshop participants will be organized in working groups, and working group sessions will alternate with plenary sessions. The conclusions and recommendations of the workshop will be considered in a plenary session on the final day. A Prospectus for the Workshop prepared by the steering committee is enclosed.

If you are able to participate in the workshop, the National Academy of Sciences will provide round-trip economy trip air fare and subsistence costs in accordance with NAS procedures. Upon your acceptance, further details will be forwarded to you. Please send the enclosed response form as soon as possible to Mary Hope Katsouros, Senior Staff Officer, Board on Ocean Science and Policy.

If you have any questions about the workshop, please call me (617/253-5937) or Mary Hope Katsouros (202/334-2714).

I look forward to your favorable reply.

Sincerely,

Carl Wunsch, Chairman

Enclosure

APPENDIX II:
LIST OF PARTICIPANTS

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Workshop on Global Observations and Understanding
of the General Circulation of the Oceans

National Academy of Sciences
Woods Hole Study Center
Woods Hole, Massachusetts

August 8-12, 1983

PARTICIPANTS

Carl I. Wunsch, Chairman
Department of Earth & Planetary
Sciences (54-1324)
Massachusetts Institute of Technology
Cambridge, MA 02139

D. James Baker, Jr.
Joint Oceanographic Institutions, Inc.
21st & Pennsylvania Avenue N.W., #316
Washington, D. C., 20006

Bert R.I. Bolin
University of Stockholm
Department of Meteorology
Arrheinius Laboratory
S-10691 Stockholm
SWEDEN

Francis P. Bretherton
National Center for Atmospheric
Research (NCAR)
P. O. Box 3000
Boulder, Colorado 80307

Wallace Broecker
Geochemistry Department
Lamont-Doherty Geological Observatory
of Columbia University
Palisades, NY 10964

Neil L. Brown
Neil Brown Instrument Systems, Inc.
P. O. Box 498
1140 Route 28A
Cataumet, Massachusetts 02534

Kirk Bryan
Geophysical Fluid Dynamics Laboratory
Princeton University
P. O. Box 308
Princeton, NJ 08540

Harry L. Bryden
Department of Physical Oceanography
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Joost A. Businger
Department of Atmospheric Research
(AK-40)
University of Washington
Seattle, Washington 98195

Anthony Calio
Assistant Administrator
National Oceanic & Atmospheric
Administration
Department of Commerce
14th and Constitution Ave., N.W. #5804
Washington, D. C., 20230

Robert E. Cheney
N/CG11, 423 Rockwall Building
National Geodetic Survey
National Oceanic & Atmospheric
Administration
6010 Executive Boulevard
Rockville, Maryland 20852

Curtis A. Collins
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

James Crease
Institute of Oceanographic Sciences
Wormley, Godalming
Surrey GU8 5UB
ENGLAND

Russ E. Davis
Scripps Institution of Oceanography
(A-030)
University of California, San Diego
La Jolla, California 92093

Robert Etkins
National Climate Program Office
6010 Executive Boulevard, #508
Rockville, Maryland 20852

Manuel E. Fiadeiro
Department of Geology and Geophysics
Yale University
P. O. Box 6666
210 Whitney Avenue
New Haven, Connecticut 06511

Nicholas P. Fofonoff
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Michael Ghil
Courant Institute - N.Y.U.
251 Mercer Street
New York, New York 10012

Dale Haidvogel
National Center for Atmospheric
Research
P. O. Box 3000
Boulder, Colorado 80307

D. E. Harrison
(54-1410)
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

Robert H. Heinmiller, Jr.
Omnet, Incorporated
70 Tonawanda Street
Boston, Massachusetts 02124

William J. Jenkins
Clark Building
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

A. D. Kirwan
Department of Marine Sciences
University of South Florida
140 - 7th Avenue South
St. Petersburg, Florida 33701

Ryuji Kimura
Ocean Research Institute
University of Tokyo
1-15-1 Minamidai
Nakano-ku TOKYO 164
JAPAN

William George Large
National Center for Atmospheric
Research (NCAR)
P. O. Box 3000
Boulder, Colorado 80307

Michel Pierre Lefebvre
Groupe deRecherches deGeodesie Spatiale
18 Avenue Edouard Belin
31055 Toulouse Cedex
FRANCE

James R. Luyten
Department of Physical Oceanography
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

James C. McWilliams
National Center for Atmospheric
Research (NCAR)
P. O. Box 3000
Boulder, Colorado 80307

James G. Marsh
Code 921
Goddard Space Flight Center
Greenbelt, Maryland 20771

Pierre Morel
World Climate Research Programme
c/o World Meteorological Organization
Case Postale No. 5
CH-1211 Geneva 20
SWITZERLAND

John Morrison
National Science Foundation
1800 G Street N.W.
Washington, D. C., 20550

Walter H. Munk
Institute of Geophysics & Planetary
Physics (A-025)
University of California, San Diego
La Jolla, California 92093

Pearn P. Niiler
Scripps Institution of Oceanography
(A-030)
University of California, San Diego
La Jolla, California 92093

Worth D. Nowlin, Jr.
Department of Oceanography
Texas A&M University
College Station, Texas 77843

William C. Patzert
Oceanic Processes Branch
National Aeronautics & Space
Administration (EE-8)
Washington, D.C., 20546

Joseph Pedlosky
Clark Laboratory, 363
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Norman A. Phillips
National Meteorological Center
National Oceanic & Atmospheric
Administration
W32 WWB Rm. 204
Washington, D.C., 20233

James F. Price
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Joseph L. Reid
Marine Life Research Group
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

Roger Revelle
Humanities Library Bldg., B-007
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

Peter B. Rhines
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Paola M. Rizzoli
Department of Meteorology & Physical
Oceanography (54-1420)
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

Allan R. Robinson
Division of Applied Sciences
Pierce Hall
Harvard University
Cambridge, Massachusetts 02038

Claes G.H. Rooth
Cooperative Institute for Marine and
Atmospheric Studies
University of Miami
4600 Rickenbacker Causeway
Miami, Florida 33149

Jorge L. Sarmiento
Geophysical Fluid Dynamics Program
Princeton University
Princeton, New Jersey 08540

Joseph Smagorinsky
21 Duffield Place
Princeton, New Jersey 08540

Robert C. Spindel
Department of Ocean Engineering
Bigelow 107
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

John Steele
Director
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Robert H. Stewart
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

Robert W. Stewart
Director-General
Marine Sciences Branch - Pacific Region
Department of the Environment
1230 Government Street, 5th Floor
Victoria, B.C., CANADA V8W 1YA

Henry Stommel
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Bertrand Thompson
Committee on Climate Changes and
the Ocean
7, Place de Fontenoy
75700 Paris
FRANCE

George Veronis
Kline Geology Laboratory
P. O. Box 6666
New Haven, Connecticut 06511

Bruce A. Warren
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Douglas C. Webb
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Ferris Webster
College of Marine Studies
University of Delaware
Lewes, Delaware 19958

Ray F. Weiss
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

W. Stanley Wilson
Oceanic Processes Branch
National Aeronautics and Space
Administration (EE-8)
Washington, D.C., 20546

John Woods
Institute fur Meereskunde
Dusternbrooker Weg 20
D-2300 KIEL
Federal Republic of Germany

National Research Council Staff

Mary Hope Katsouros
Nancy G. Maynard
Debra Luks
Doris Taylor-Lucas
Sam Amos
Tom O'Neill, BASC

APPENDIX III:
BACKGROUND PAPERS

NOTE: The background papers are printed here as received from the authors, not only in the interest of expediting the publication of this report, but also because the committee felt there would be value to the reader in having the papers reproduced as facsimiles of the originals. These papers have not undergone the usual NRC review process. In some instances, the authors will submit final versions of the papers for journal publication.

BACKGROUND PAPERS APPEAR IN ALPHABETICAL SEQUENCE BY AUTHOR

Broecker, Wallace S., 1983. Measurement Strategies for Nutrient Constituents

Bryan, Kirk, 1983. Climate Models Related to Stream 3: Outstanding Technical Problems

Bryan, Kirk, 1983. A Generic Heat Flux Experiment

Davis, R.E., 1983. Drifters in the Study of Oceanic General Circulation

Haidvogel, D.B., W.R. Holland, 1983. General Circulation Modeling

Heinmiller, Robert, 1983. Physical Oceanographic Instrumentation Developments for a World Ocean Circulation Experiment

Large, W.G., 1983. Air-Sea Transfer Processes: The Surface Observations

Ledwell, James R., 1983. Purposeful Tracer Experiments

McWilliams, James C., 1983. A Concept of WOCE

Munk, Walter, Carl I. Wunsch, 1983. Ocean Acoustic Tomography - Basin Scale Applications

Nowlin, W.D., Jr., Joseph L. Reid, 1983. A WOCE Density Program

O'Brien, James J., 1983. Satellite Wind Observations

Phillips, Norman A., 1983. Data Analysis and Modelling

Sarmiento, J.L., 1983. Tracers and Modeling

Weiss, R. F., 1983. Chemical Tracer Measurements

Woods, J.D., 1983. Seasonal Variation and Water Mass Conversion

Wunsch, Carl I., 1983. A World Ocean Circulation Experiment

Wunsch, Carl I., 1983. Sea Surface Topography -- Toward a WOCE Strategy

WOCE PLANNING DOCUMENT

Measurement Strategies for Nutrient Constituents

Wallace S. Broecker

Prepared for WOCE Workshop

in Woods Hole

August, 1983

INTRODUCTION

A combined knowledge of the distributions of the concentrations of the nutrient species O_2, NO_3, PO_4, ΣCO_2 and alkalinity and of the distributions of the rates of respiration (for soft tissue) and of dissolution (for $CaCO_3$ and opal) would provide very powerful constraints on the patterns and rates of mixing in the sea. These constraints will be especially important for those waters which will not receive substantial inputs of anthropogenic tracers over the next few decades. We already have excellent information regarding the distribution of these nutrient properties in the sea. Given the geographic location, the depth, the potential temperature and the salinity of a water sample the nutrient constituent concentrations could be interpolated from existing data sets with an accuracy about as good as that obtained by routine nutrient measurement programs*. What is now needed is a program dedicated to mapping the respiration and dissolution funtions. The combination of flux measured using sediment traps and benthic chambers offers a means to do this.

NUTRIENT-ORGANISM INTERACTIONS

The distributions of O_2, NO_3, PO_4, ΣCO_2, alkalinity and H_4SiO_4 within the sea are determined by the interaction between the movement of water through the sea and by the formation and destruction of the soft and hard tissues of organisms. For O_2, NO_3 and PO_4 the differences are entirely the result of the photosynthesis-respiration cycle. Soft tissues are created by plants in the surface water of the sea and consumed by animals and bacteria living at all depths in the sea and on the sea floor. The photosynthesis-respiration cycle leads to a steady state depletion in NO_3 and PO_4

---

*Exceptions would be high latitude surface waters and a few places in the deep sea.

in the sea's surface waters and a corresponding enrichment at depth. O_2 is near saturation in surface waters and is depleted in deep waters. The distribution of H_4SiO_4 is related entirely to its utilization by diatoms and radiolarians to form their opaline tests coupled with inorganic dissolution of these tests within the sea and on the sea floor.Most of the opal formed in surface waters is returned to solution in the deep waters. The alkalinity of seawater is altered in two ways, by the formation and dissolution of $CaCO_3$ and by the utilization and regeneration of HNO_3 as part of the photosynthesis-respiration cycle. The ΣCO_2 content of seawater is altered by the formation and destruction of both $CaCO_3$ and soft tissue. It is also being gradually increased as anthropogenic CO_2 invades the sea. These interactions are summarized in table 1.

DERIVED PROPERTIES

In order to understand how the distributions of these constituents of sea salt relate to the cycles of soft tissue, $CaCO_3$, and opal, it is necessary to convert the measured concentrations to derived properties.

1) The concentrations of ΣCO_2 and alkalinity must be normalized to a constant salinity (35.00°/oo). This normalization is important for these two properties because their concentration ranges in the sea are small compared to their mean values. Thus, salinity differences account for a significant part of their variability in the sea.

2) The contribution of nitrate to alkalinity is removed as follows:

$$ALK^C = ALK^N + (NO_3-16)$$

where ALK^N is the salinity normalized alkalinity. The variations in this nitrate-corrected and salinity-normalized alkalinity (i.e. ALK^C) should then be entirely the result of $CaCO_3$ formation and dissolution.

3) The contribution of $CaCO_3$ formation and dissolution is removed from the ΣCO_2 distribution as follows:

$$\Sigma CO_2^* = \Sigma CO_2^N - \frac{(ALK^C - ALK^{STD})}{2}$$

Table 1. Property-organism interaction summary

	Soft tissue	$CaCO_3$	Opal
O_2	Yes	No	No
NO_3	Yes	No	No
PO_4	Yes	No	No
ΣCO_2	Yes	Yes	No
Alk	Yes*	Yes	No
H_4SiO_4	No	No	Yes

*Via HNO_3 uptake and release.

Table 2. Derived property definitions.

$$AOU = O_2^{Sat} - O_2^{Obs}.$$

$$ALK^C = ALK^{Obs}\left(\frac{35.0}{S}\right) + (NO_3^{Obs} - 16) = ALK^N + (NO_3^{Obs} - 16)$$

$$\Sigma CO_2^* = \Sigma CO_2^{Obs}\left(\frac{35.0}{S}\right) + \frac{(ALK^C - 2318)}{2} = \Sigma CO_2^N + \frac{(ALK^C - 2318)}{2}$$

$$PO_4^o = PO_4^{Obs} - \left.\frac{P}{-O_2}\right)_{Org} AOU = PO_4^{Obs} - \frac{AOU}{175}$$

$$\Sigma CO_2^{*o} = \Sigma CO_2^* - \left.\frac{C}{O_2}\right)_{Org} AOU = \Sigma CO_2^* - \frac{AOU}{1.34}$$

where ALK^C is as defined above, ALK^{STD} is an arbitrarily chosen reference value (i.e., 2318 μeq/kg) and ΣCO_2^N is the salinity normalized ΣCO_2 concentration. It must be kept in mind that ΣCO_2^* includes the anthropogenic effects as well as the photosynthesis-respiration effects.

4) O_2 is converted to AOU (apparent oxygen utilization) by subtracting the observed O_2 concentration from the saturation value (at the potential temperature of the water). This procedure removes the temperature dependence of the O_2 content and focuses attention on the amount consumed during respiration.

In this format we have indicators of photosynthesis-respiration (AOU), of opal formation-dissolution (H_4SiO_4) and of calcium carbonate formation-dissolution (ALK^C).

The definitions are summarized in table 2. Although the distributions of NO_3, PO_4 and ΣCO_2^* are all controlled by the photosynthesis-respiration cycle, they are not redundant to AOU or to one another. This lack of redundancy stems from the different surface boundary conditions for the four properties. AOU is near zero everywhere at the sea surface. The concentrations of NO_3 and PO_4 are near zero in all open ocean surface waters warmer than 16°C (except along the equator of the Pacific Ocean where strong upwelling occurs). Both properties have non-zero values in surface waters colder than 16°C. The distribution of nitrate in these waters is shown in Figure 1. Indeed deep waters colder than 16°C all show finite preformed (i.e. initial) concentrations of NO_3 and PO_4. These preformed values are obtained by multiplying the mean O_2 utilization to PO_4 and NO_3 generation ratios for marine organic residues by the observed AOU value and subtracting this quantity from the observed PO_4 and NO_3 values

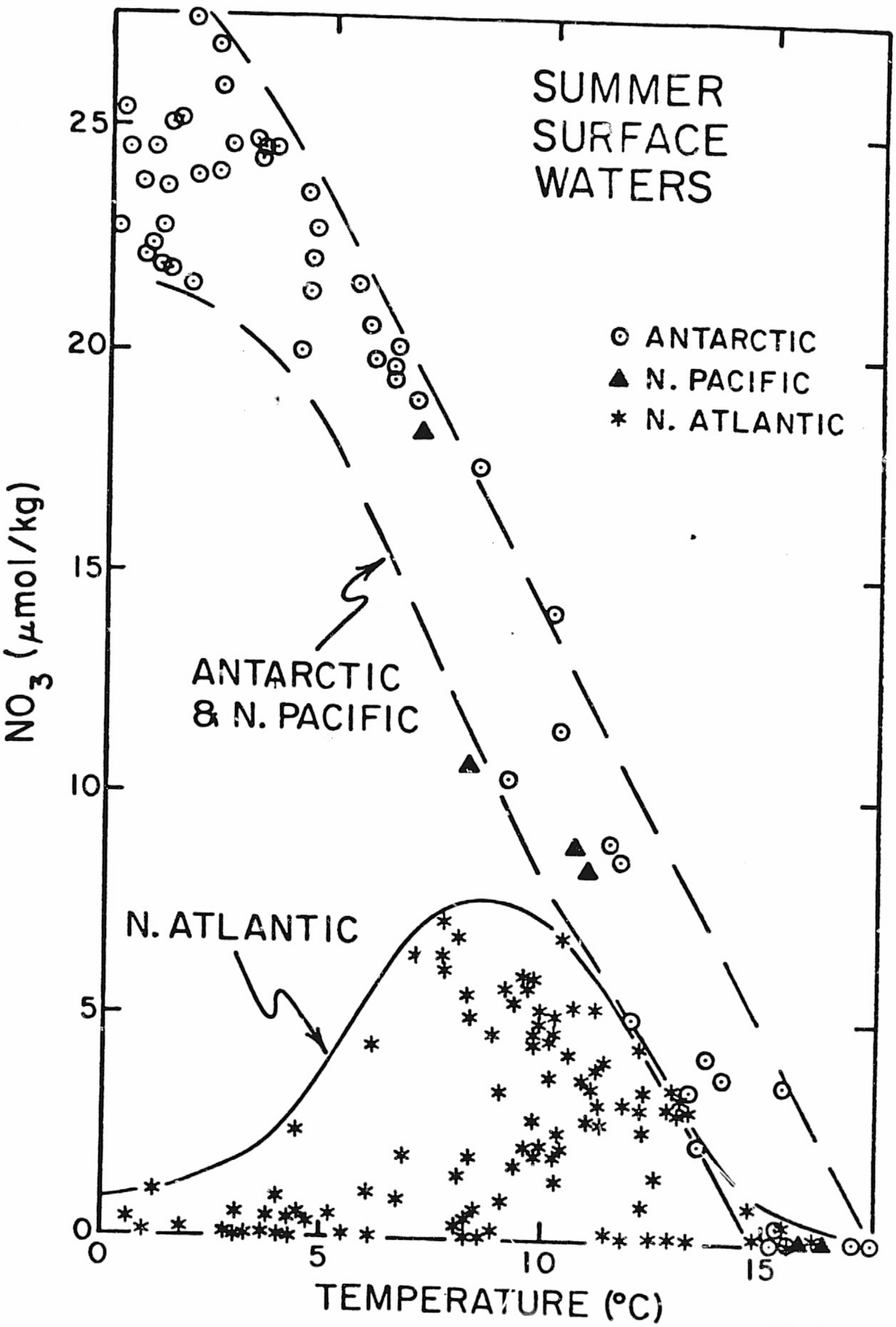

SUMMER
SURFACE
WATERS
ANTARCTIC
N. PACIFIC
N. ATLANTIC
ANTARCTIC
& N. PACIFIC
N. ATLANTIC
NO_3 (μmol/kg)
TEMPERATURE (°C)
0
5
10
15
20
25
0
5
10
15

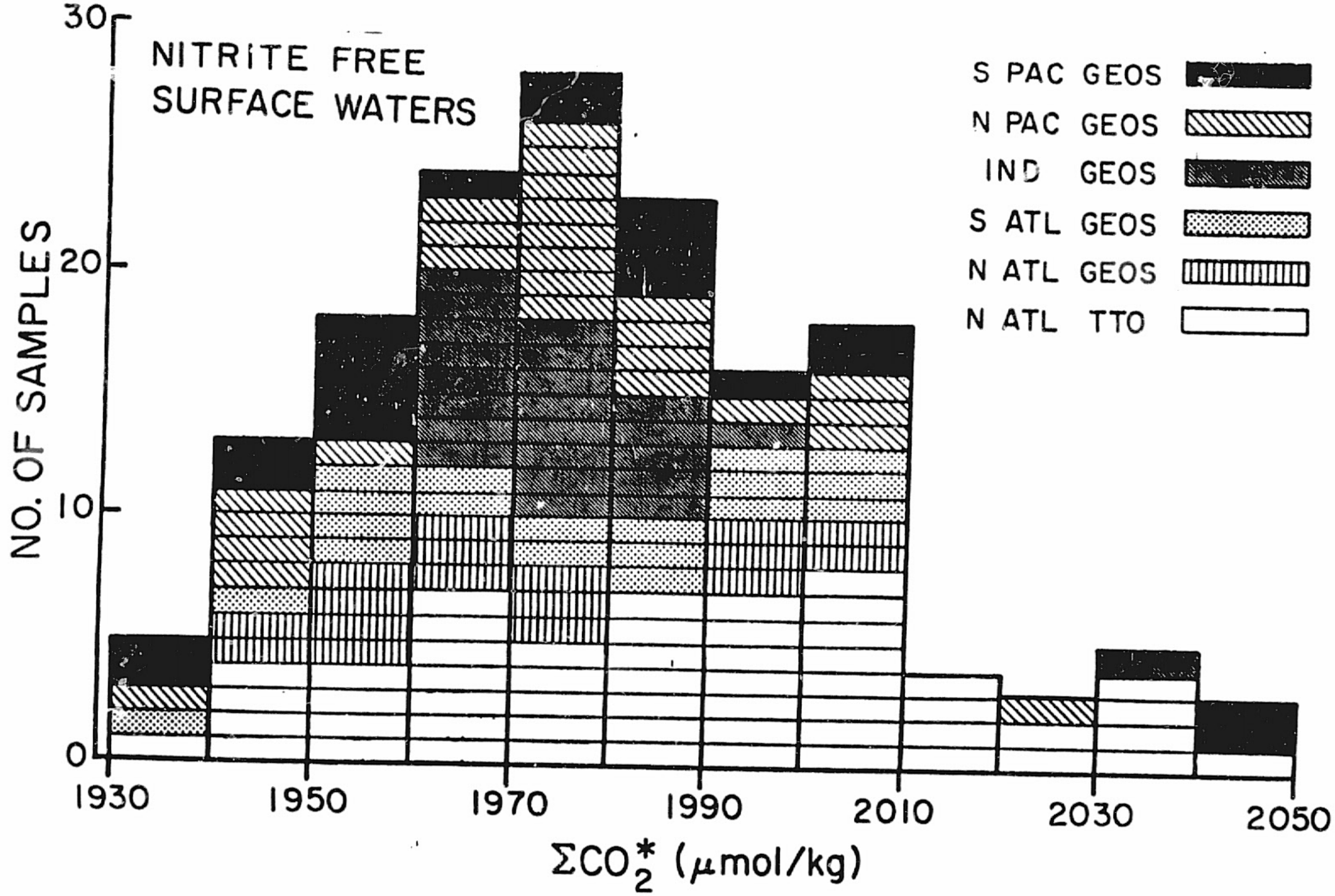
FIG. 1

(see table 2). The ΣCO_2^* values are non-zero throughout the sea. As shown in Figure 2, surface waters warmer than 16°C show a narrow range ΣCO_2^* concentration (standard deviation about 1.5%).

As shown in Figure 3, the distributions of nitrate and silicate in surface waters are similar (i.e., warm waters are free of both while the cold Antarctic surface waters are rich in both). It should be noted, however, that the high nitrate values extend to lower latitudes than do the high silicate values. As we shall see, the deviations from this base value for cold surface waters are closely correlated with the NO_3 and PO_4 concentrations. It must be emphasized that this correlation is not to be expected as CO_2 can be transported through the atmosphere from one region of the sea surface to another while NO_3 and PO_4 cannot. Also, ΣCO_2^* contains a significant anthropogenic component while NO_3 and PO_4 do not.

DISTRIBUTION PATTERNS FOR AOU, H_4SiO_4, AND ALKALINITY

Profiles of AOU, ALK^C, and H_4SiO_4 are shown for five stations in the ocean in Figures 4 through 8. Also shown in these figures are the distributions of preformed phosphate, i.e. PO_4^*, with depth. The latter property allows deep waters of Antarctic (high PO_4^*) and of northern Atlantic (low PO_4^*) to be distinguished. Maps showing the trends of AOU and H_4SiO_4 at 4 km depth are shown in Figure 9. Sections of AOU and H_4SiO_4 are compared in Figures 10, 11, and 12. As shown is that the maximum H_4SiO_4 concentration is found at greater depths than the maximum AOU concentration. This difference must be related to shallower mean depth for respiration than for opal dissolution. For deep ocean sites this could be taken to indicate that organic matter residues are largely eaten as they fall through the water column while opal line tests largely fall to the bottom before



NITRITE FREE
SURFACE WATERS
S PAC GEOS
N PAC GEOS
IND GEOS
S ATL GEOS
N ATL GEOS
N ATL TTO
NO. OF SAMPLES
0
10
20
30
1930
1950
1970
1990
2010
2030
2050
ΣCO_2^* (μmol/kg)

FIG. 2

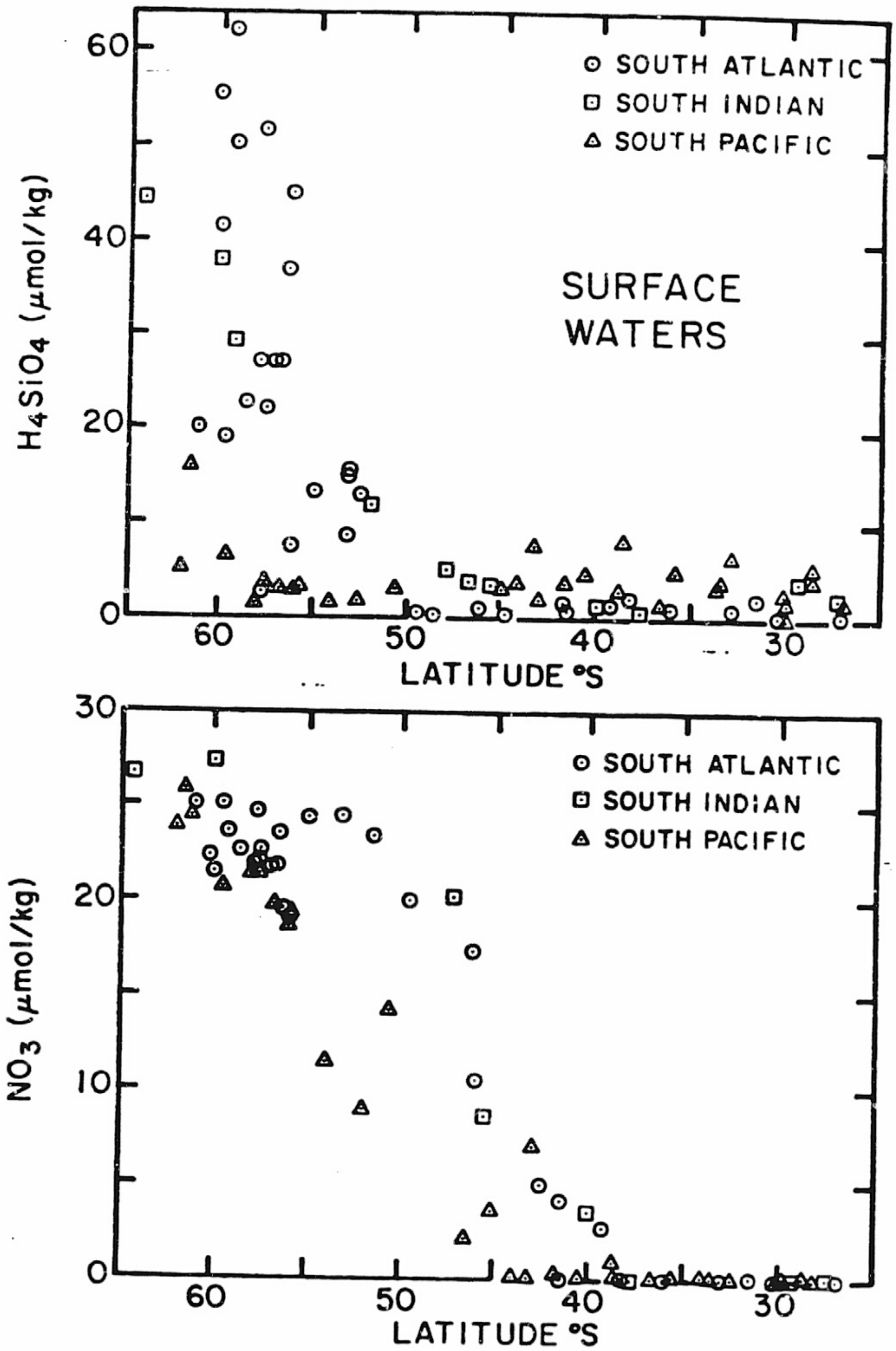
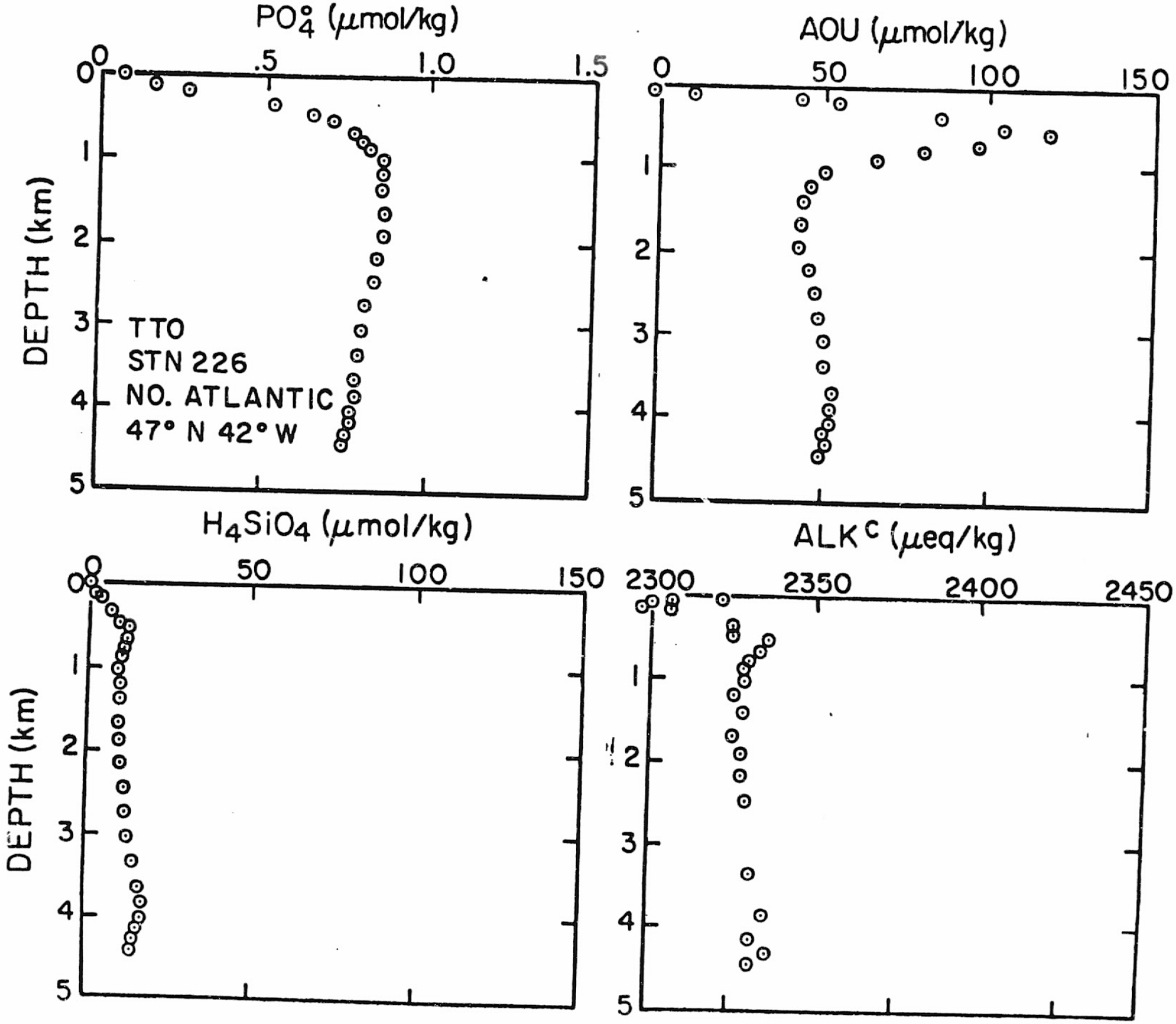

60
40
20
0
H4SiO4 (μmol/kg)
SOUTH ATLANTIC
SOUTH INDIAN
SOUTH PACIFIC
SURFACE
WATERS
60
50
40
30
LATITUDE °S
30
20
10
0
NO3 (μmol/kg)
SOUTH ATLANTIC
SOUTH INDIAN
SOUTH PACIFIC
60
50
40
30
LATITUDE °S

FIG. 3



PO$_4^\circ$ (μmol/kg)
0
.5
1.0
1.5
AOU (μmol/kg)
0
50
100
150
DEPTH (km)
0
1
2
3
4
5
TTO
STN 226
NO. ATLANTIC
47° N 42° W
H$_4$SiO$_4$ (μmol/kg)
ALK^C (μeq/kg)
2300
2350
2400
2450

FIG. 4

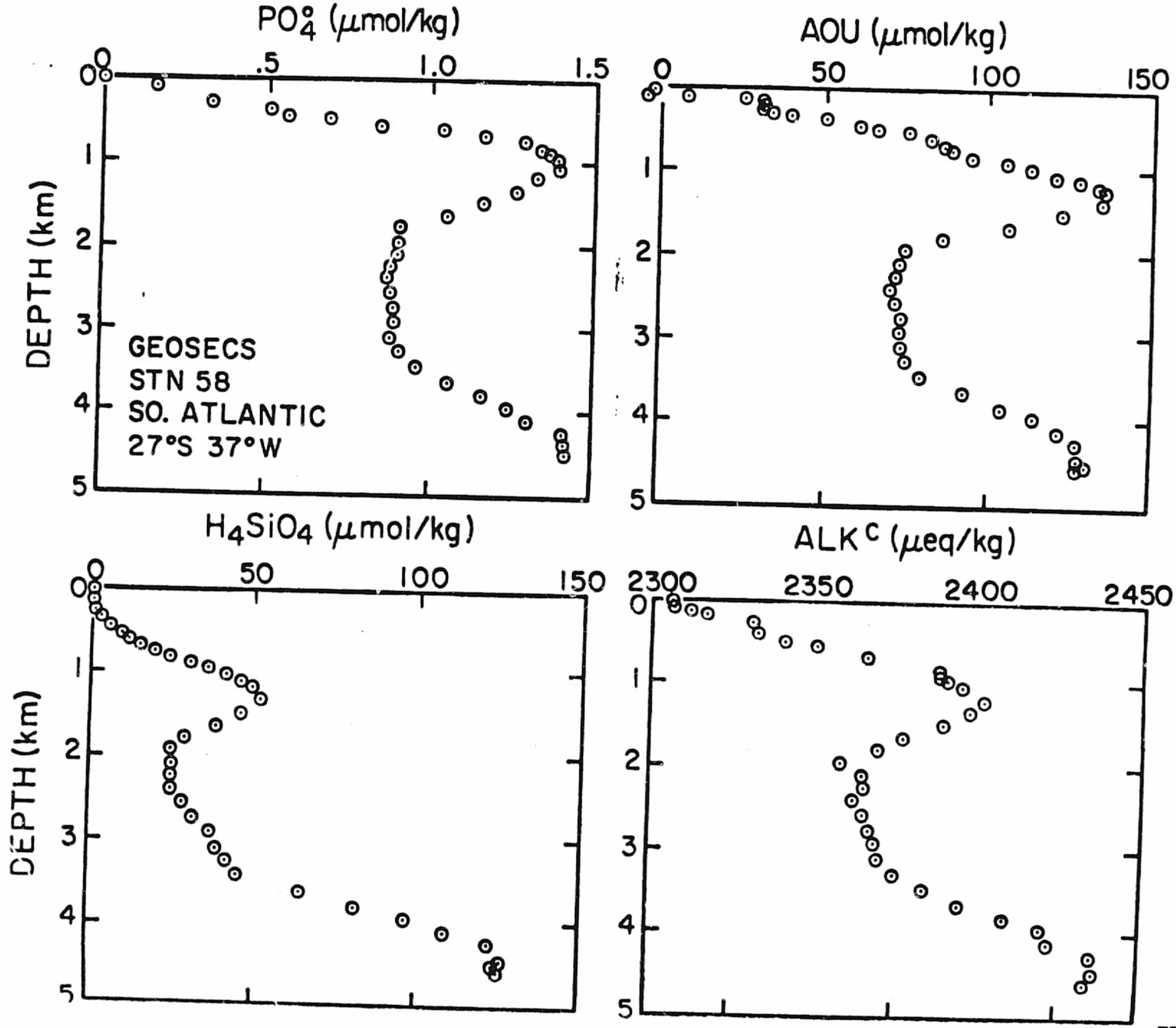
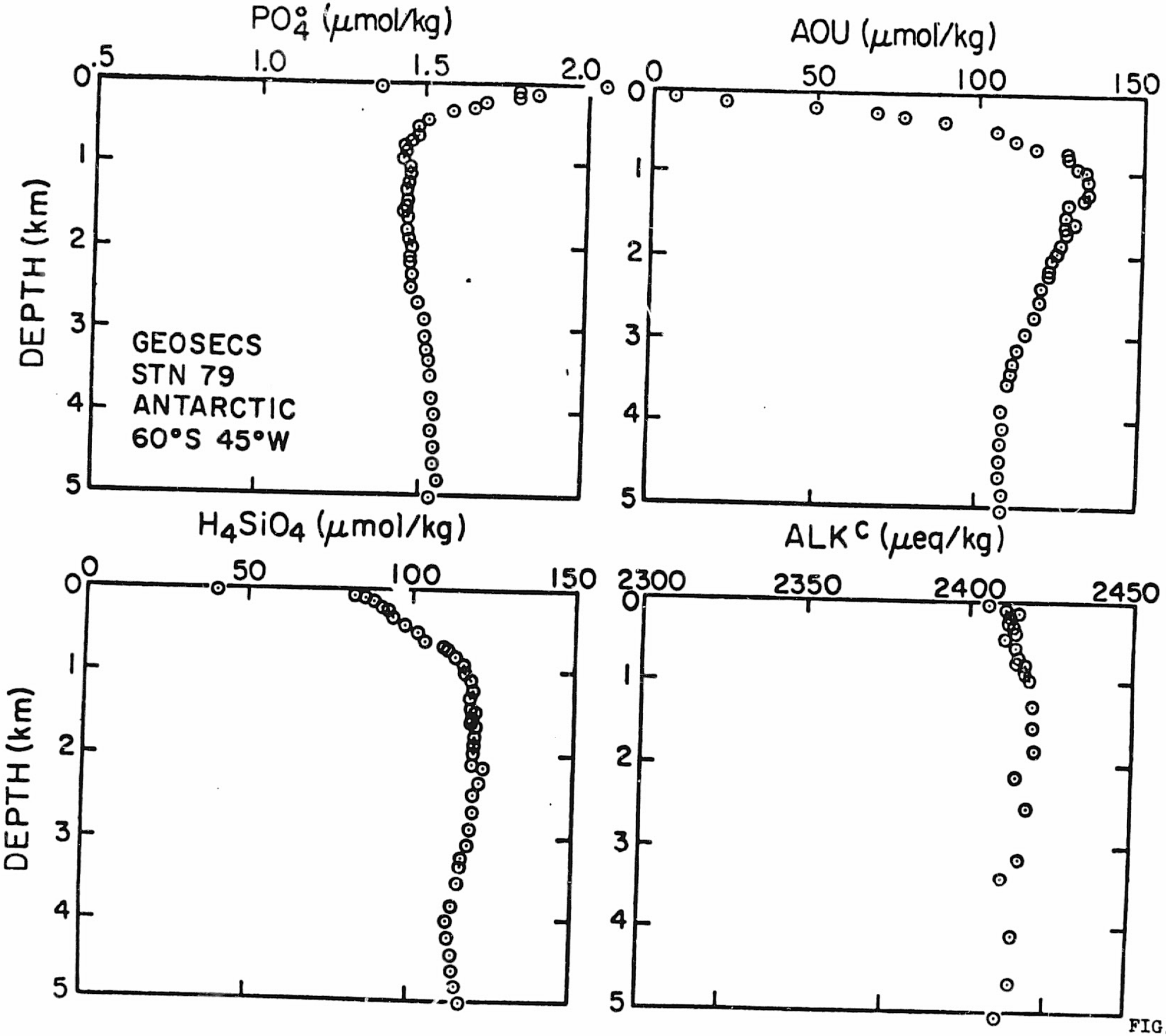
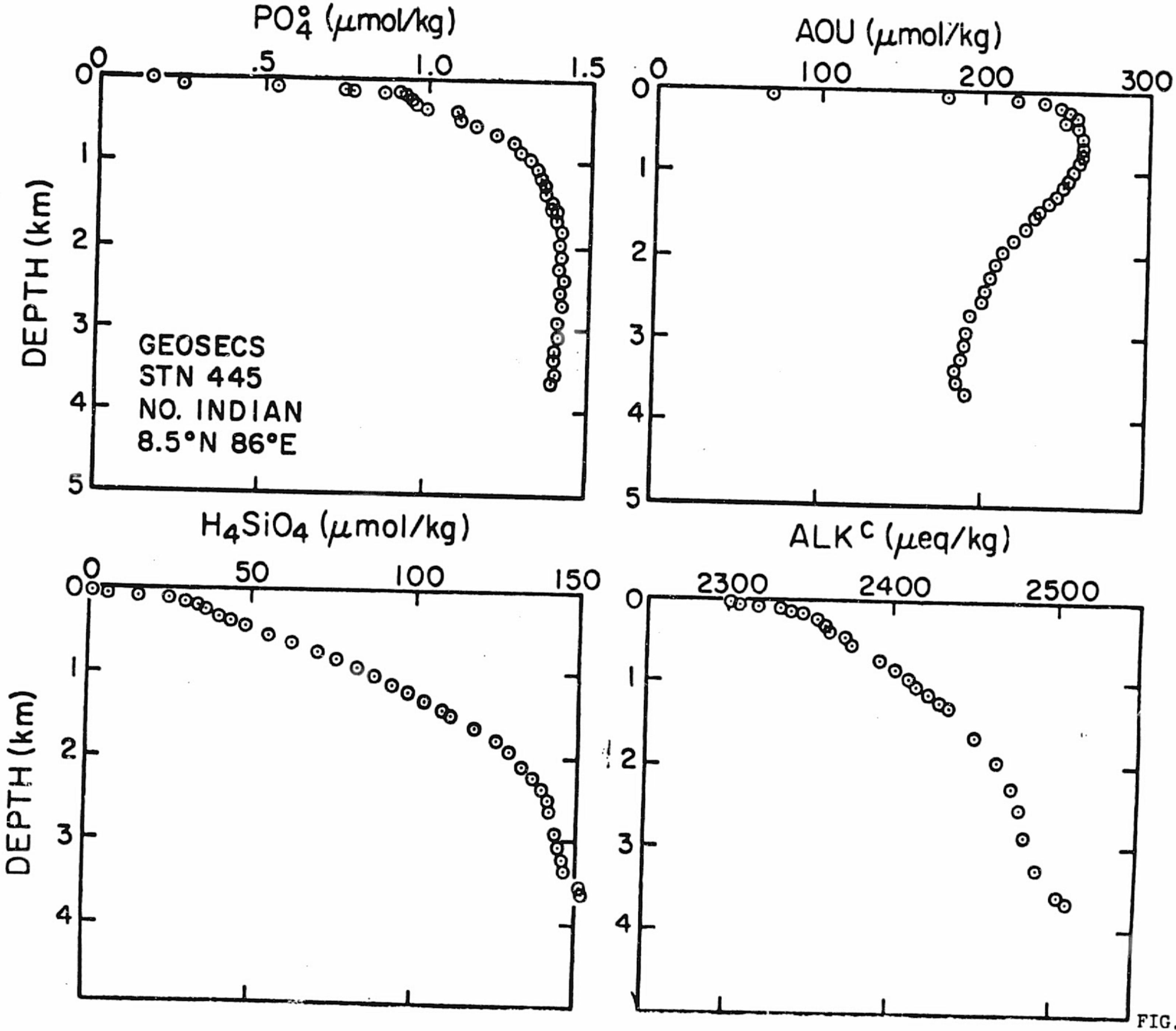

PO°4 (μmol/kg)
0
.5
1.0
1.5
DEPTH (km)
0
1
2
3
4
5
GEOSECS
STN 58
SO. ATLANTIC
27°S 37°W
AOU (μmol/kg)
0
50
100
150
H4SiO4 (μmol/kg)
0
50
100
150
ALK C (μeq/kg)
2300
2350
2400
2450

FIG. 5



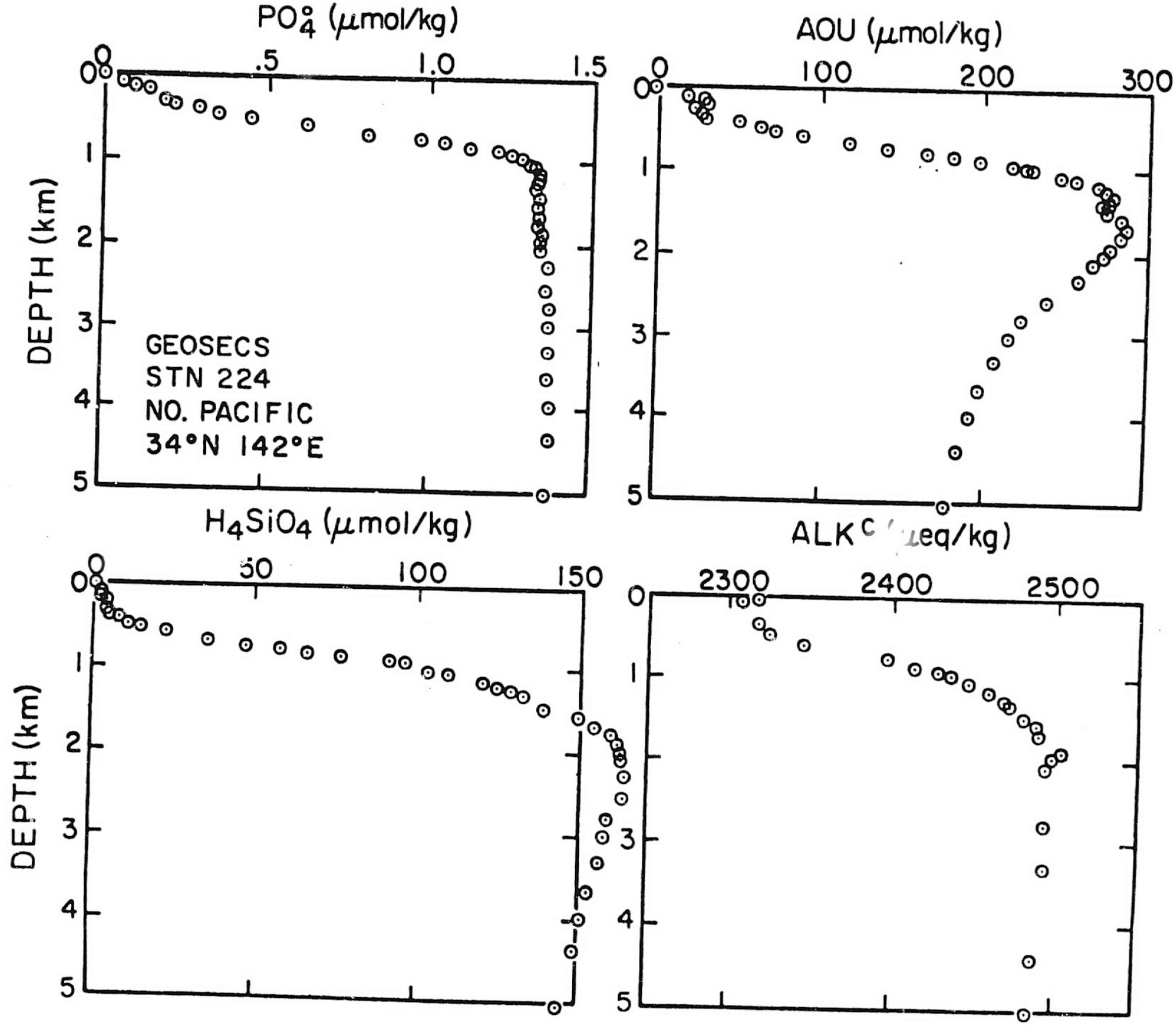
PO°4 (μmol/kg)
0.5
1.0
1.5
2.0
0
1
2
3
4
5
DEPTH (km)
GEOSECS
STN 79
ANTARCTIC
60°S 45°W
AOU (μmol/kg)
0
50
100
150
H4SiO4 (μmol/kg)
ALK C (μeq/kg)
2300
2350
2400
2450

FIG. 6



PO$_4^\circ$ (μmol/kg)
0
.5
1.0
1.5
DEPTH (km)
0
1
2
3
4
5
GEOSECS
STN 445
NO. INDIAN
8.5°N 86°E
AOU (μmol/kg)
0
100
200
300
H$_4$SiO$_4$ (μmol/kg)
0
50
100
150
ALK^C (μeq/kg)
2300
2400
2500

FIG. 1


PO$_4^\circ$ (μmol/kg)
0
.5
1.0
1.5
AOU (μmol/kg)
0
100
200
300
DEPTH (km)
0
1
2
3
4
5
GEOSECS
STN 224
NO. PACIFIC
34°N 142°E
H$_4$SiO$_4$ (μmol/kg)
0
50
100
150
ALK^c (μeq/kg)
2300
2400
2500

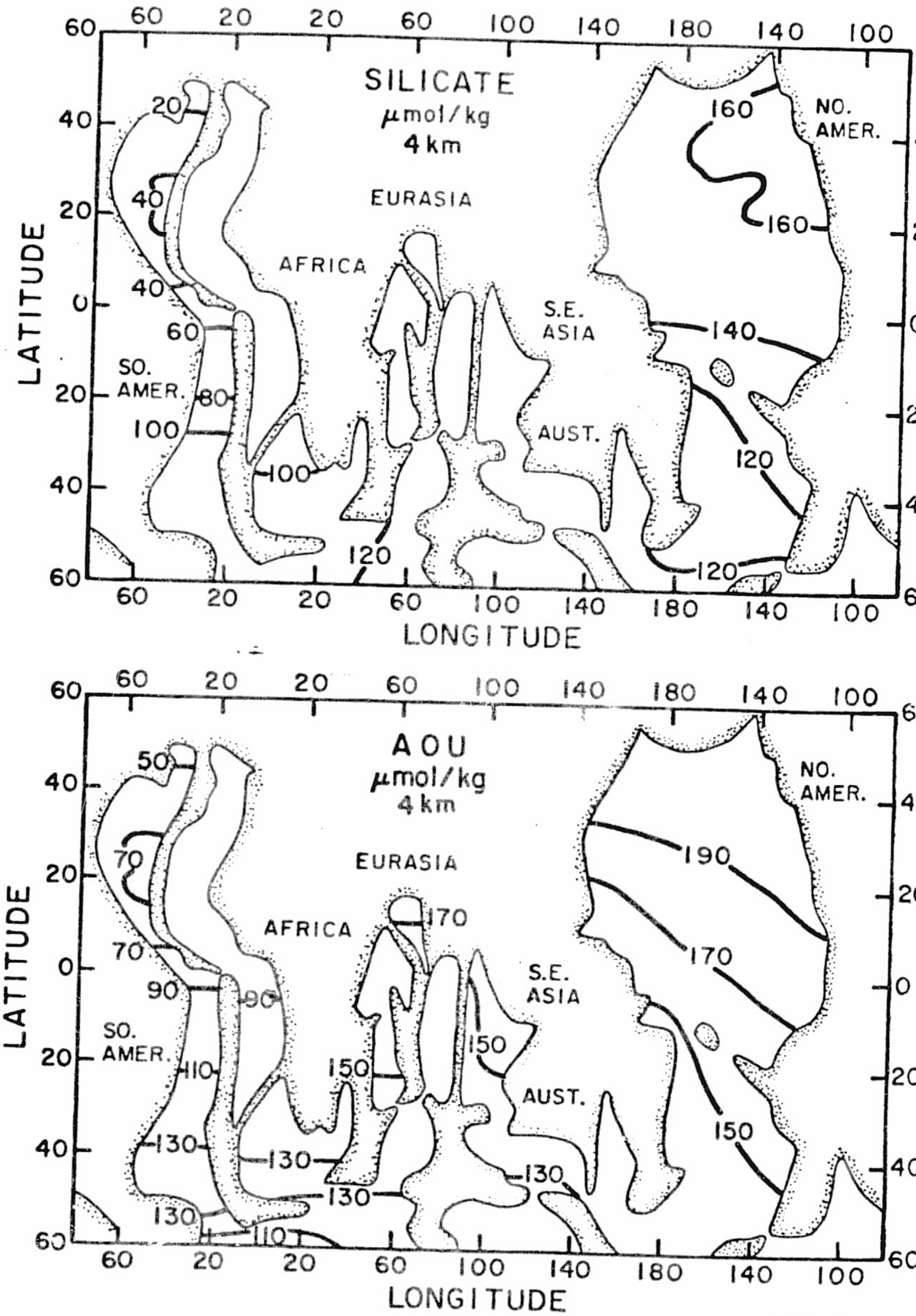

SILICATE
µmol/kg
4 km
EURASIA
AFRICA
S.E. ASIA
AUST.
SO. AMER.
NO. AMER.
20
40
40
60
80
100
100
120
120
120
140
160
160
LATITUDE
LONGITUDE
AOU
µmol/kg
4 km
EURASIA
AFRICA
S.E. ASIA
AUST.
SO. AMER.
NO. AMER.
50
70
70
90
90
110
130
130
130
130
130
110
150
150
150
170
170
190
LATITUDE
LONGITUDE

FIG. 9

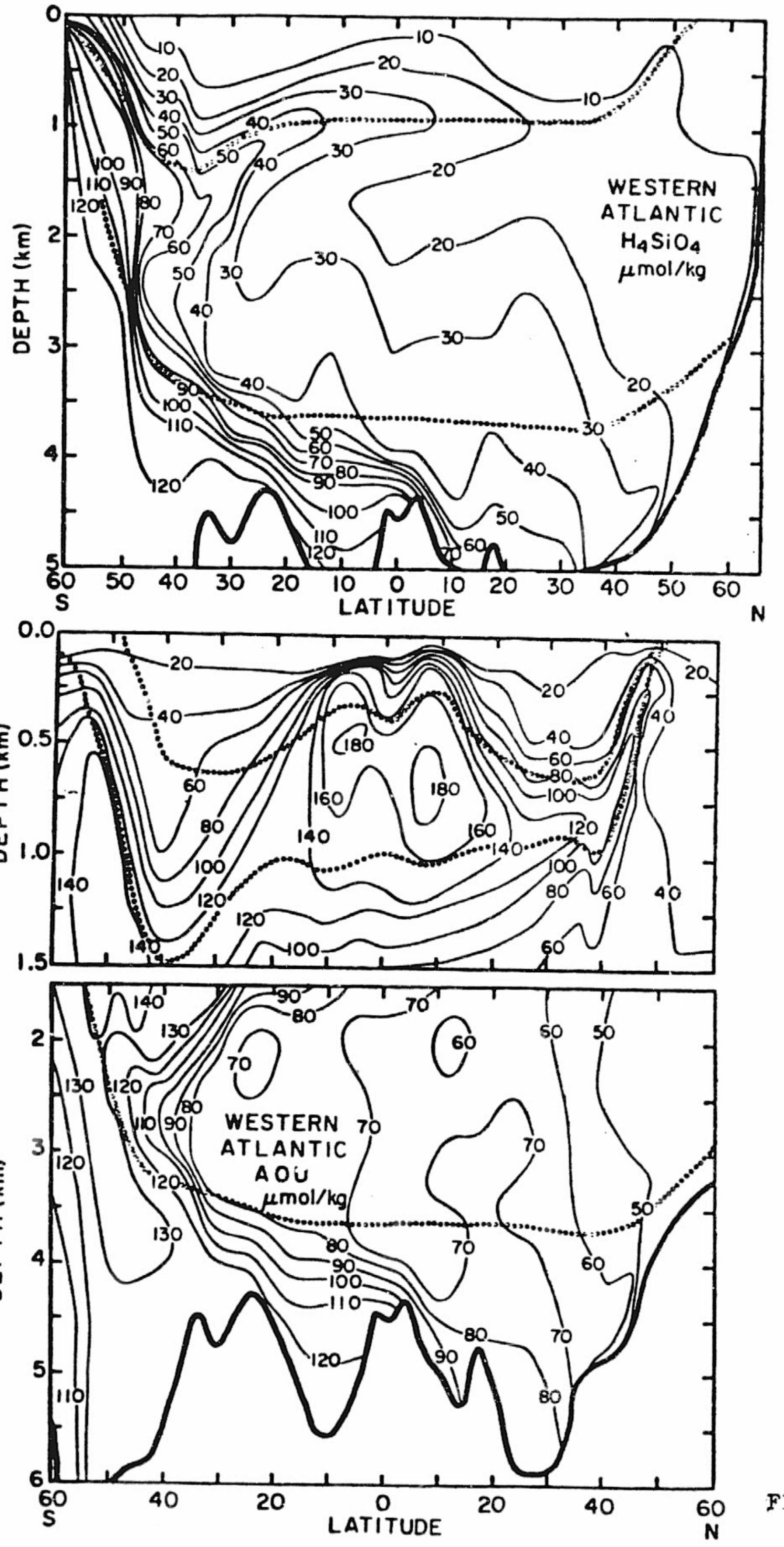

WESTERN
ATLANTIC
H4SiO4
µmol/kg
DEPTH (km)
LATITUDE
S
N
DEPTH (km)
WESTERN
ATLANTIC
AOU
µmol/kg
DEPTH (km)
LATITUDE
S
N

FIG. 10

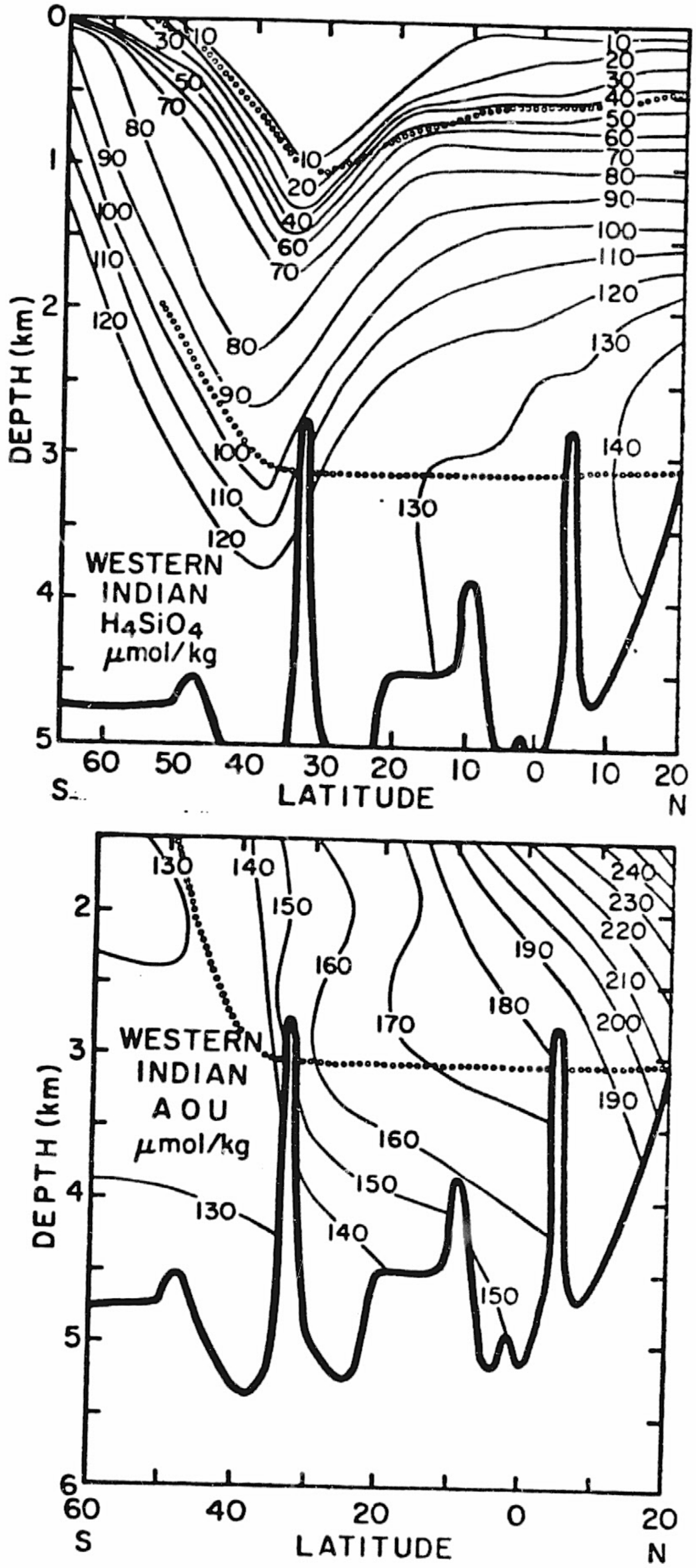

WESTERN
INDIAN
H_4SiO_4
μmol/kg
DEPTH (km)
LATITUDE
S
N
WESTERN
INDIAN
AOU
μmol/kg
DEPTH (km)
LATITUDE
S
N

FIG. 11

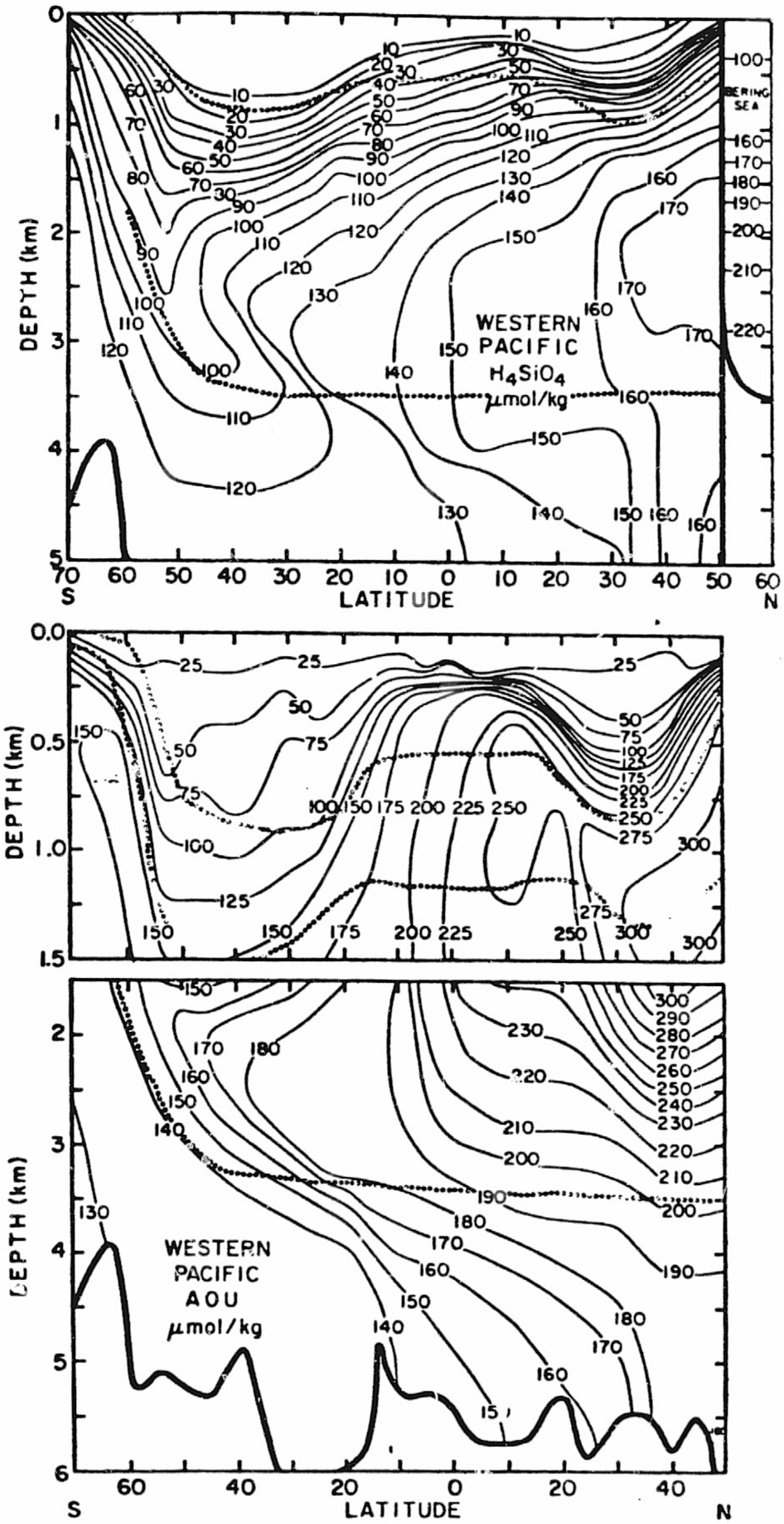
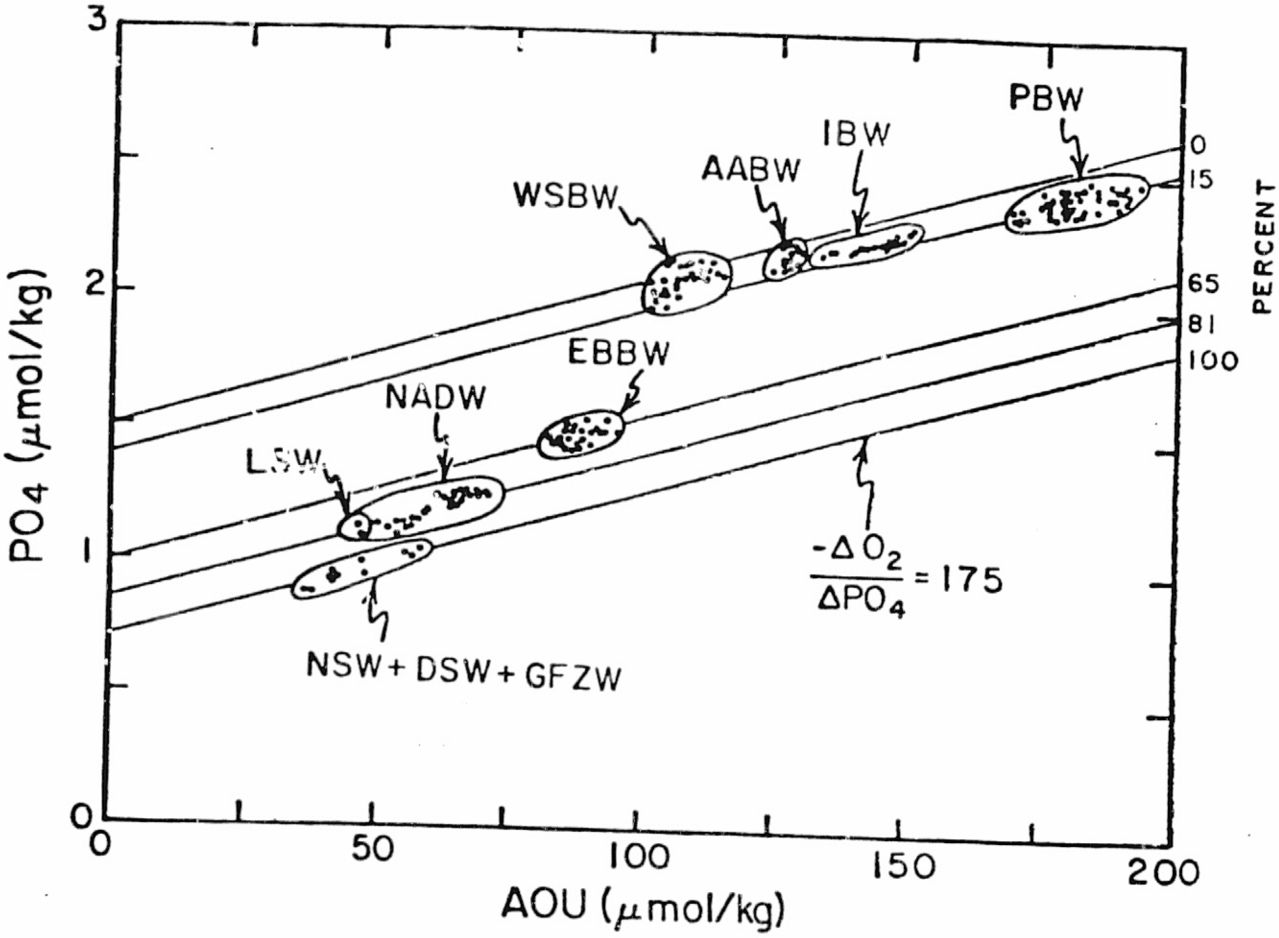

WESTERN
PACIFIC
H_4SiO_4
μmol/kg
BERING SEA
DEPTH (km)
LATITUDE
S
N
WESTERN
PACIFIC
AOU
μmol/kg
DEPTH (km)
LATITUDE
S
N

FIG. 12

dissolving. On the other hand, it may reflect a large difference in the ratio of soft tissue residue production to opal production between sites underlain by the abyssal ocean and sites underlain by the shallow margins of the ocean.

The plot of phosphate vs. AOU in Figure 13 demonstrates how the difference between the surface water pattern for these two properties can be used as a water mass tracer. Shown in this diagram are the major deep water types in today's ocean. As can be seen for a given deep water type the points line up along lines with a slope of approximately 175 to 1. This ratio is the same as that obtained from chemical variations along isopycnals in the sea (both in the thermocline and deep sea). If indeed further studies show that these so-called Redfield ratios are similar from area to area and depth to depth in the sea then they can be used to yield preformed PO_4 values for deep waters. As can be seen in Figure 13, all the deep water types lie between the trends for those waters thought by oceanographers to be the two major deep water sources (i.e. Weddell Sea bottom water in the Antarctic, Norwegian and Labrador Sea deep waters in the northern Atlantic). If the situation is so simple then the relative contributions of the two sources to any given deep water can be read from this diagram.

As suggested above, the distributions of NO_3, PO_4, and ΣCO_2C in the sea are highly correlated. This is shown in Figures 14 and 15. The slopes of these trends are consistent with the so-called Redfield ratios derived from analyses along isopycnal horizons (see Table 3 for summary). It should be kept in mind that the ΣCO_2^*-PO_4 relationship has been influenced by the invasion of anthropogenic CO_2. The possible impact of this



WSBW Weddell Sea Bottom Water
AABW Antarctic Bottom Water (South Atlantic)
IBW Indian Ocean Bottom Water
PBW Pacific Ocean Bottom Water
EBBW Atlantic Ocean Eastern Basin Bottom Water
NADW North Atlantic Deep Water
LSW Labrador Seas Deep Water
DSW Denmark Straits Overflow Water
NSW Norwegian Sea Deep Water

FIG. 13

Table 3. Redfield ratio estimates based on the analysis of chemical gradients along isopycnal horizons in the sea (Broecker et al. in press)

	P	N	C	$-O_2$
ATLANTIC (thermocline)	1	17.8	130	177
INDIAN (thermocline)	1	14.5	131	170
INDIAN (deep)	1	13.8	125	164
PACIFIC (deep)	1	14.5	131	170

Table 4. Equilibrium anthropogenic CO_2 contents of surface waters.

	ATM P_{CO_2} 10^{-6} ATM	ATM ΔP_{CO_2} 10^{-6} ATM	Excess ΣCO_2 µm/kg	
			25°C	2°C
Preanthropogenic	265*	0	0	0
GEOSECS 1972-1973	328	63	53	35
TTO North Atlantic 1981	340	75	63	42

*Based on Bern ice core measurements.

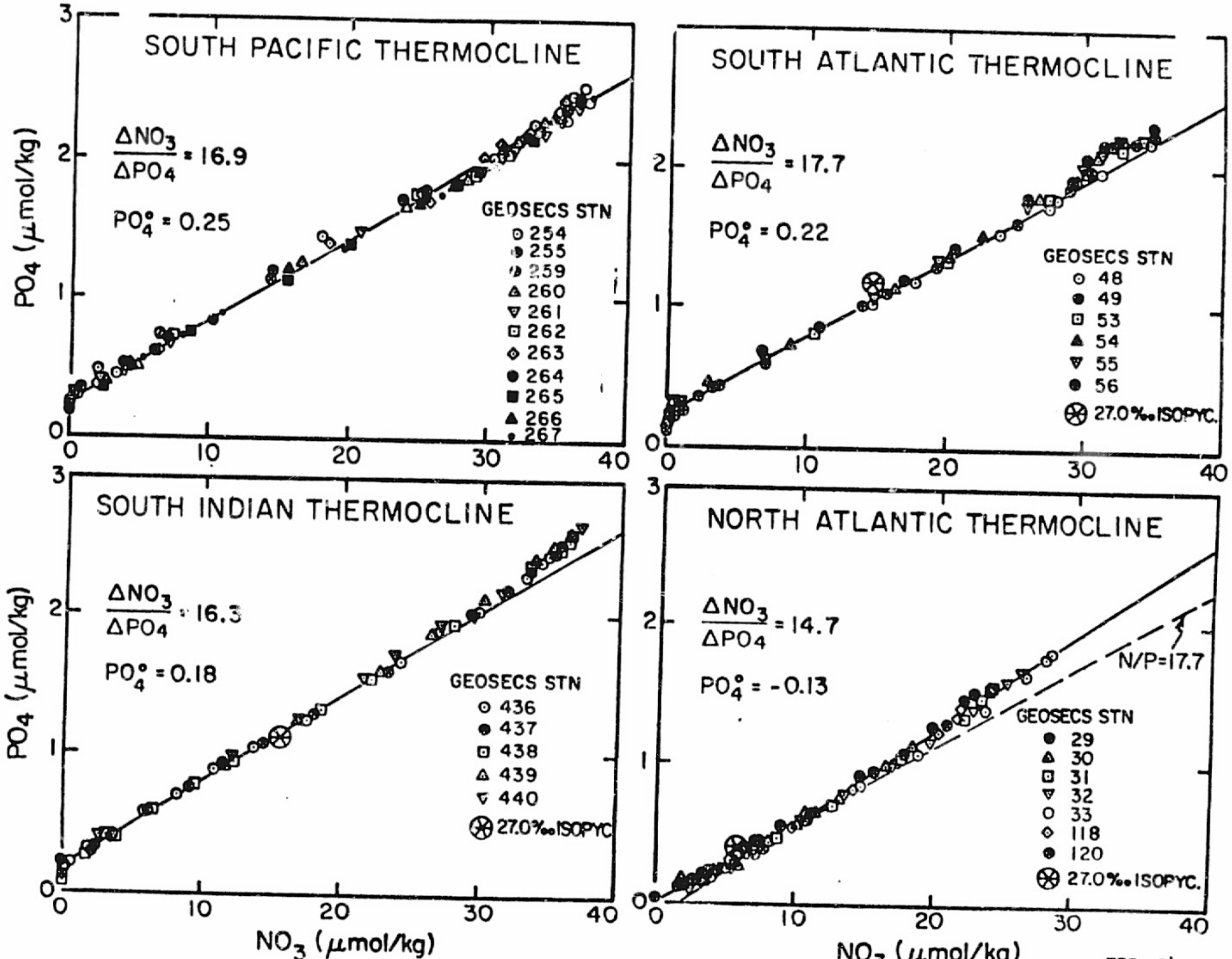

SOUTH PACIFIC THERMOCLINE
ΔNO3 / ΔPO4 = 16.9
PO4° = 0.25
GEOSECS STN
254
255
259
260
261
262
263
264
265
266
267
SOUTH ATLANTIC THERMOCLINE
ΔNO3 / ΔPO4 = 17.7
PO4° = 0.22
GEOSECS STN
48
49
53
54
55
56
27.0‰ ISOPYC.
SOUTH INDIAN THERMOCLINE
ΔNO3 / ΔPO4 = 16.3
PO4° = 0.18
GEOSECS STN
436
437
438
439
440
27.0‰ ISOPYC
NORTH ATLANTIC THERMOCLINE
ΔNO3 / ΔPO4 = 14.7
PO4° = -0.13
N/P=17.7
GEOSECS STN
29
30
31
32
33
118
120
27.0‰ ISOPYC.
PO4 (μmol/kg)
NO3 (μmol/kg)

FIG. 14

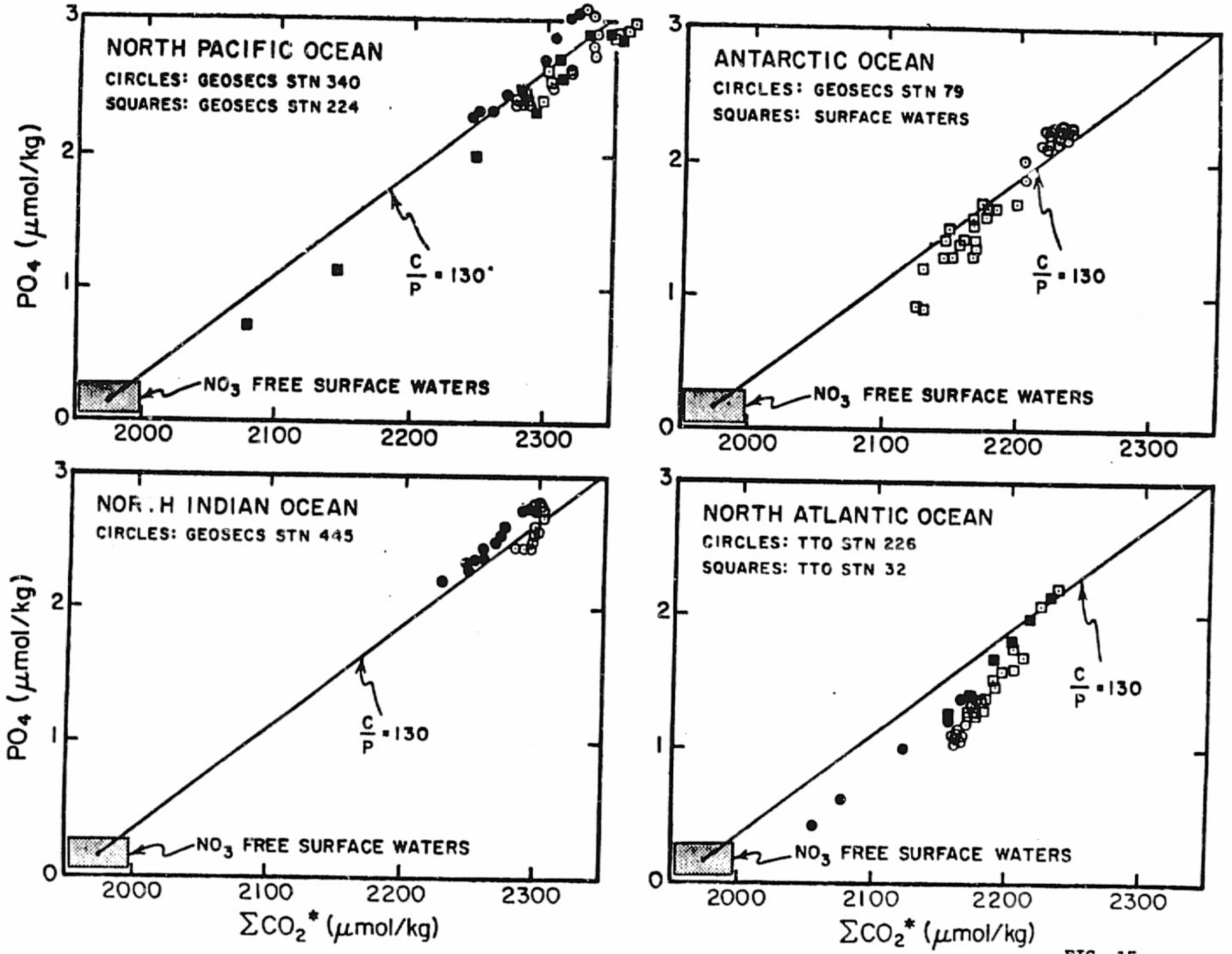

NORTH PACIFIC OCEAN
CIRCLES: GEOSECS STN 340
SQUARES: GEOSECS STN 224
C/P = 130
NO3 FREE SURFACE WATERS
ANTARCTIC OCEAN
CIRCLES: GEOSECS STN 79
SQUARES: SURFACE WATERS
C/P = 130
NO3 FREE SURFACE WATERS
NORTH INDIAN OCEAN
CIRCLES: GEOSECS STN 445
C/P = 130
NO3 FREE SURFACE WATERS
NORTH ATLANTIC OCEAN
CIRCLES: TTO STN 226
SQUARES: TTO STN 32
C/P = 130
NO3 FREE SURFACE WATERS
PO4 (μmol/kg)
ΣCO2* (μmol/kg)

FIG. 15

invasion is summarized in Table 4. Since the anthropogenic impact on low PO_4 waters will on the average have been larger than that on high PO_4 waters, the slope ($\Sigma CO_2^*/PO_4$) must at the time of the GEOSECS and TTO surveys have been smaller than when the ocean was pristine (see Figure 16). Unfortunately we currently have no means to reliably separate the anthropogenic and natural contributions to the ocean's ΣCO_2 distribution.

While the correlations between NO_3 and PO_4 and between ΣCO_2^* and PO_4 are generally quite good they are not perfect. Waters originating in the northern Atlantic have higher ΣCO_2^* and NO_3 values for any given PO_4 content than do those originating in the southern ocean. Both offsets likely owe their origin to transport through the air from one region of the ocean to the other. This is easily understood for CO_2 which surely flows from regions of the surface ocean with higher P_{CO_2} values to regions with lower P_{CO_2} values. For nitrogen this transport is via N_2 which is being converted to NO_3 by organisms living mainly in surface waters and produced from NO_3 by bacteria living in anaerobic sediments and in the anaerobic waters found in the thermoclines of the eastern Pacific and northwestern Indian Oceans. The imbalance in the net for these two processes (production of NO_3 is likely greater than destruction in the Atlantic and destruction is likely higher than production in the Indian and Pacific Oceans) is likely counteracted by a flow of N_2 out of the Pacific and Indian Oceans through the atmosphere into the Atlantic. Because the P_{N_2} differences in surface waters necessary to negotiate the required transport lie well within the uncertainty of P_{N_2} measurements there is as yet no way to directly assess rate at which this process operates.

In light of the extensive survey of the nutrient distributed with in the ocean as part of the GEOSECS program, the question arises as to what more needs to be done. In my estimation, the most important thing to be

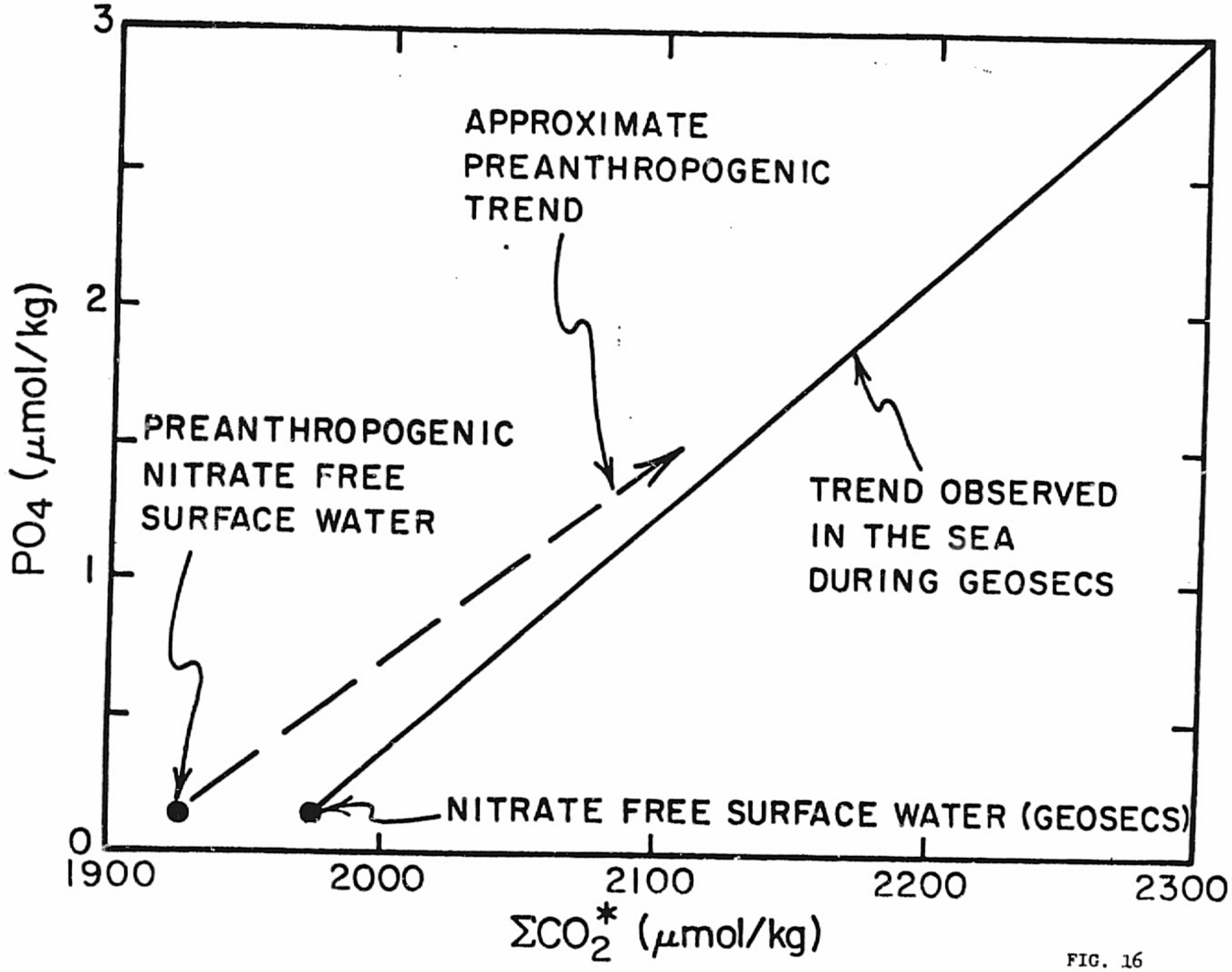

APPROXIMATE
PREANTHROPOGENIC
TREND
PREANTHROPOGENIC
NITRATE FREE
SURFACE WATER
TREND OBSERVED
IN THE SEA
DURING GEOSECS
NITRATE FREE SURFACE WATER (GEOSECS)
PO_4 (μmol/kg)
ΣCO_2^* (μmol/kg)
0
1
2
3
1900
2000
2100
2200
2300

FIG. 16

done is to carry out programs aimed at determining the rates of respiration, of calcium carbonate dissolution, and of opal dissolution. However, before discussing this aspect of nutrient geochemistry let me complete the discussion of water column distribution studies by saying what should be done to tidy up our knowledge.

As can be seen from the map of the GEOSECS stations (Figure 17) the greatest gaps in our coverage lie in the northeastern and southeastern Pacific Ocean. The TTO Program has improved the coverage in the north and equatorial Atlantic and a proposal is pending to carry out a similar detailed survey in the south Atlantic. Thus, with the exceptions mentioned above, to the extent that the nutrient properties of the ocean are at steady state, the data from the TTO and GEOSECS Programs provide excellent global coverage.

What should WOCE do? My suggestion is that WOCE on its θ-S tracks designed to determine the density sections emphasize accurate O_2 and H_4SiO_4 measurements. These properties show the highest sensitivity (i.e. signal to measurement error ratio). WOCE should also support the continuation of TTO expeditions, on which highly accurate measurements of all six of the nutrient properties are made (i.e., O_2, H_4SiO_4, PO_4, NO_3, ΣCO_2 and ALK).

One of the objectives of the TTO program is to obtain evidence which will allow the excess ΣCO_2 of anthropogenic origin to be determined. To do this requires data of the highest possible precision and accuracy. If, as some biologists have suggested the life cycles in the sea have been perturbed, then deviations from steady state for O_2, NO_3 and PO_4 may also be

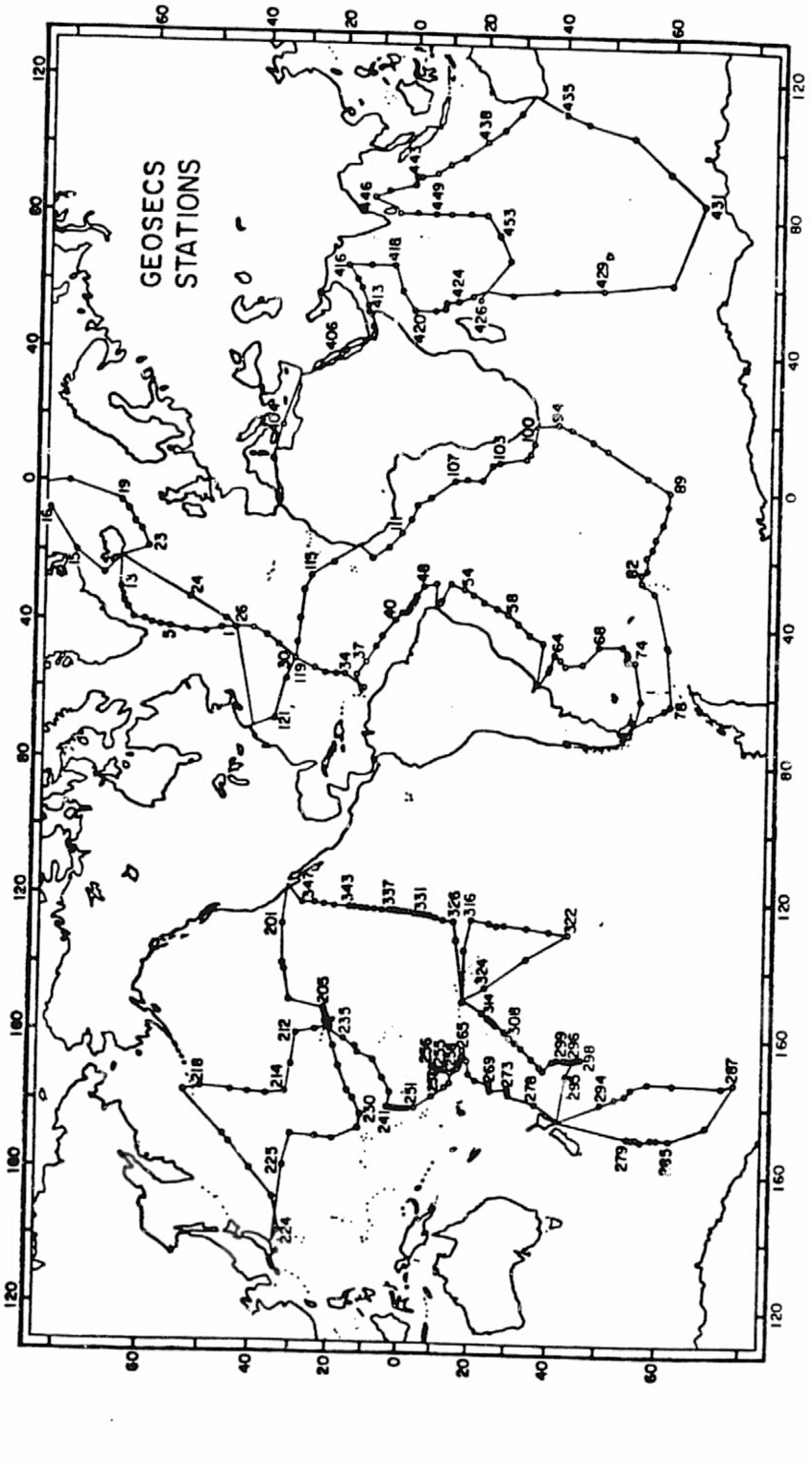


FIG. 17

found. If these are to be spotted, data of the highest possible accuracy and precision must be achieved. As these studies will in all likelihood not be compatible with WOCE density section program they will have to be carried out on dedicated TTO type expeditions.

RESPIRATION AND DISSOLUTION FUNCTIONS

To date the distributions of nutrient constituents in the ocean have been interpreted mainly in a qualitative manner. If these interpretations are to be quantified we need to learn far more than we now know about the distribution of the rates of respiration, $CaCO_3$ dissolution and opal dissolution as a function of location in the water column and location on the sea floor. We have a few facts which allow some broad generalizations.

1) The substances of interest are recycled in the sea many times. The ratio residence time in the sea to ocean ventilation time is about 100 for PO_4, 10 or more for NO_3, about 180 for ΣCO_2 and about 1000 for Ca. Because of this the distributions of these constituents do not depend on their points of entry to or removal from the sea. Rather, it depends entirely on the interaction between the patterns of water movement and patterns of production and degradation of the biogenic components of interest.

2) Over 99% of the organic matter generated in the sea is recycled. Over 95% of the opal generated in the sea is recycled. Over 95% of the $CaCO_3$ generated above areas of the sea floor lying beneath the lysocline is recycled. Because of this the rain rates of these substances as measured in sediment traps provides a measure of the combined rate at which they are destroyed in the water column and sediment beneath the trap. For areas where the sea floor projects above the lysocline, the calcite accumulation rate as measured by radiocarbon dating of the sediments must be subtracted from the rain rate measured in the traps.

3) The opal content of marine sediments shows large variations. It is large under areas of upwelling and quite low elsewhere (see Figure 18). This distribution almost certainly reflects inhomogeneities in the production rate of opal with geographic location. (i.e., production is high in areas of upwelling). Because the sea is everywhere strongly undersaturated with respect to opal, where the the opal content of sediments is high the dissolution rate will also be high. Thus, from the distribution of opal contents in recent marine sediments we get a first order picture of the pattern of H_4SiO_4 regeneration i.e., the map in figure 18 can be taken as a first order indication as to the pattern of H_4SiO_4 input into the deep sea.

4) From radiocarbon dating of deep sea sediments collected above the lysocline we get the idea that the rain rate of calcite must be reasonably uniform over the open ocean. The pattern of calcite content in open ocean sediments has to do mainly with the elevation of the sea floor with respect to the lysocline. Sediments well below the lysocline have lost virtually all their calcite to dissolution. Those from near the lysocline have lost part of their calcite to dissolution. Those from above the lysocline have lost only a little of their calcite to dissolution. Thus the sediment-water interface component of $CaCO_3$ dissolution is confined largely to the abyssal ocean. How much comes from dissolution in the water column (due perhaps to ingestion by organisms is not known). The map in figure 18 shows the boundary between $CaCO_3$ rich and $CaCO_3$ poor sediments. From this map and our knowledge of the rates of $CaCO_3$ accumulation above the lysocline a rough estimate of the input resulting from sea floor dissolution of $CaCO_3$ can be made.

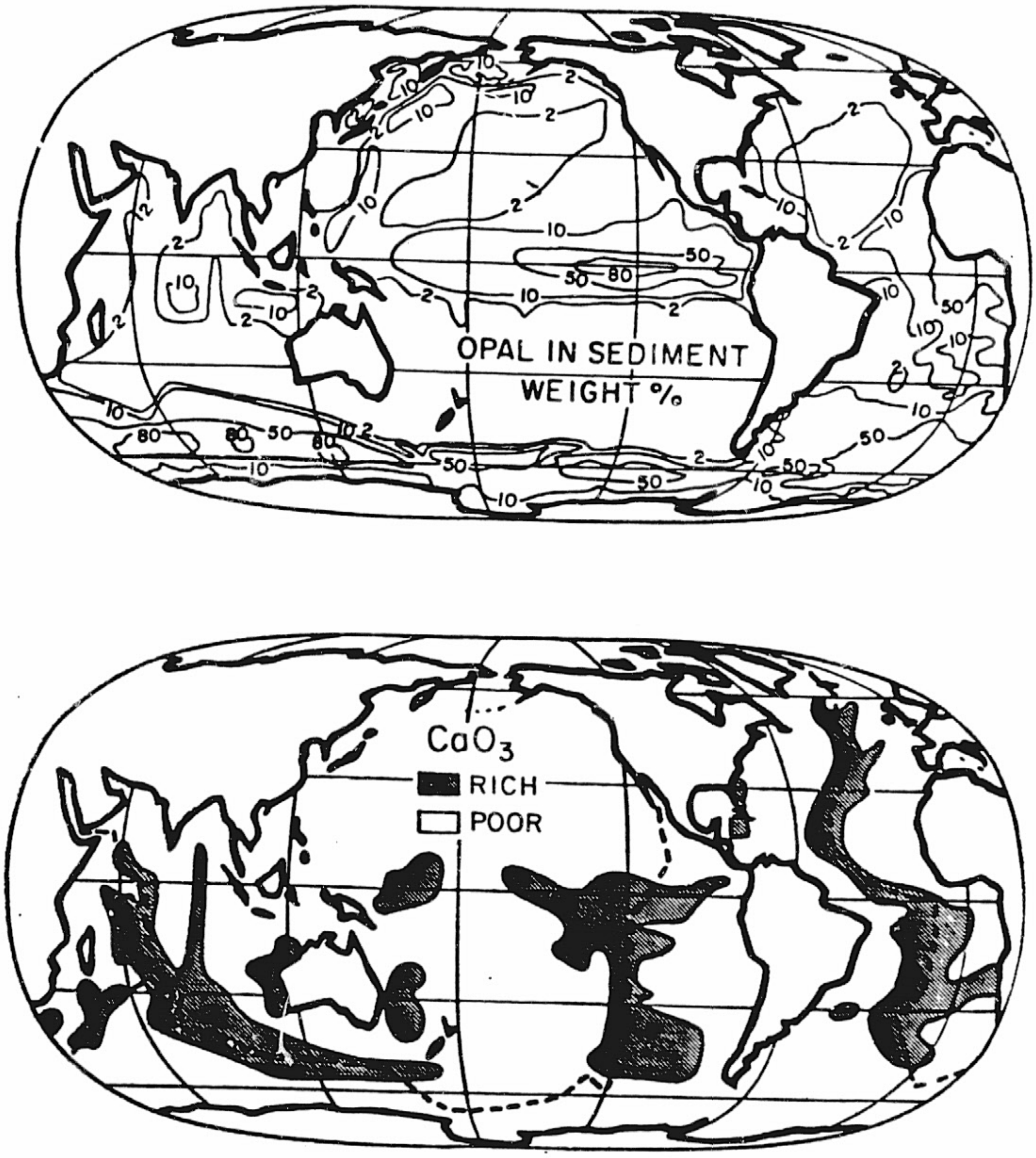


FIG. 18

5) The consensus among marine biologists is that the rate of soft tissue production is considerably higher in areas of upwelling and in coastal environments than in the open ocean (see figure 19). If so, there must be a gradient in the respiration rate away from basin margins. The shallow bottom depths in these areas insure that this extra respiration will be confined to shallow to intermediate depths (i.e., to the thermocline rather than the deep sea). To what extent respiration along the margins contributes to the strong O_2 minima seen in the thermocline of the basin interior has yet to be determined.

6) From the excesses in ALK^C and H_4SiO_4 and the deficiencies in O_2 in deep waters we know that the <u>total</u> amounts of opal, $CaCO_3$, and organic tissue (as measured in moles of Si, C, and Ca respectively) falling to the deep sea must be similar in magnitude. The radiocarbon based rate of deep sea ventilation tells us that the <u>mean</u> flux for each must be about 0.4 moles/m^2 yr. These averages help us to put the fluxes measured at specific places on the sea floor into context.

While these generalizations allow a qualitative picture of the rates of respiration, $CaCO_3$ dissolution and opal dissolution to be constructed they do not provide the quantitative input functions needed if we are to take advantage of the information carried by the water column nutrient distributions. A prime objective of WOCE should therefore be to improve our knowledge of these input functions!

We have two relatively new tools which can be used to this end; the sediment trap and the benthic flux chamber. While both techniques are still being improved and verified, projecting to the start of WOCE both should be ready for routine use.

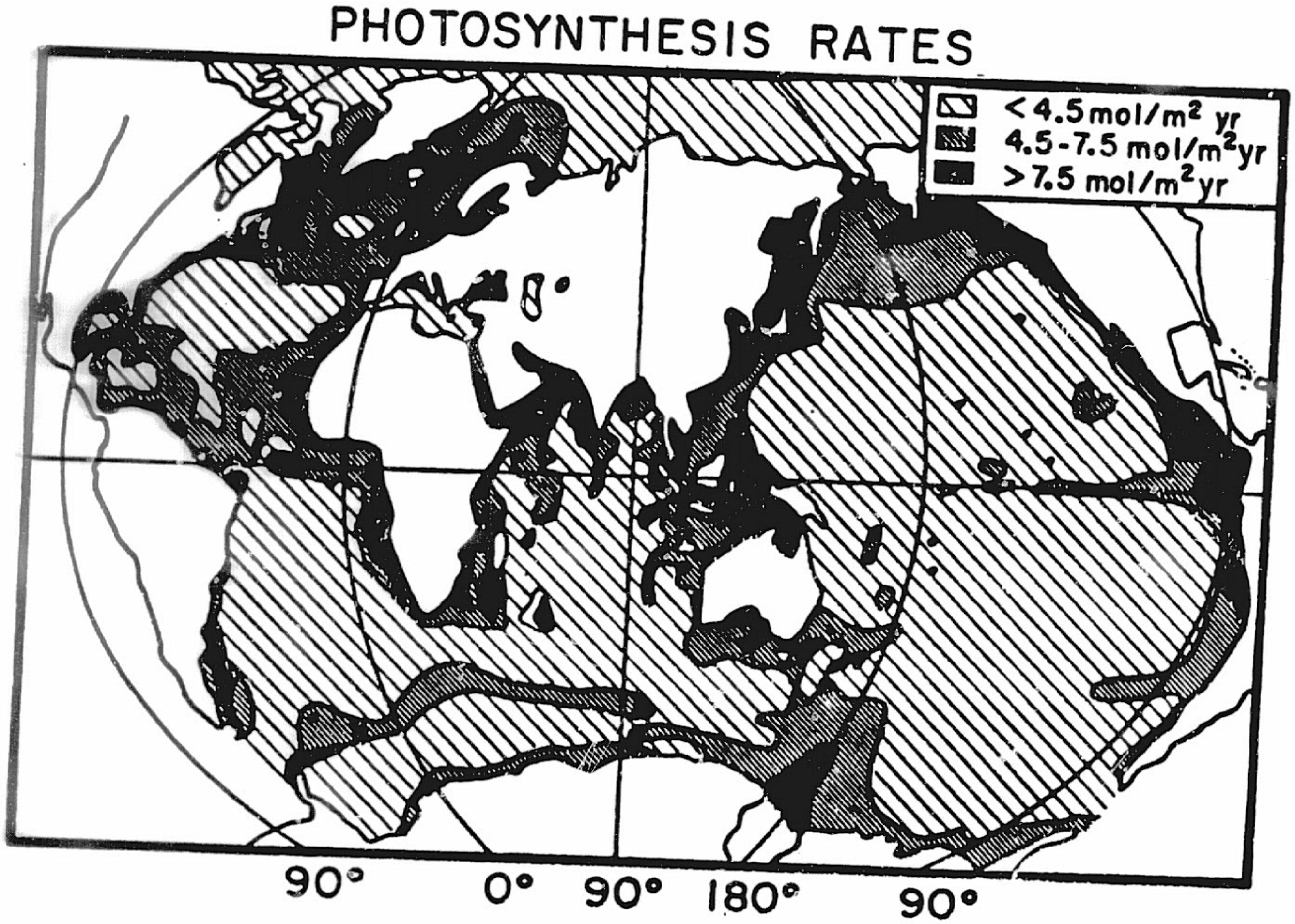

PHOTOSYNTHESIS RATES
< 4.5 mol/m² yr
4.5-7.5 mol/m² yr
> 7.5 mol/m² yr
90°
0°
90°
180°
90°

FIG. 19

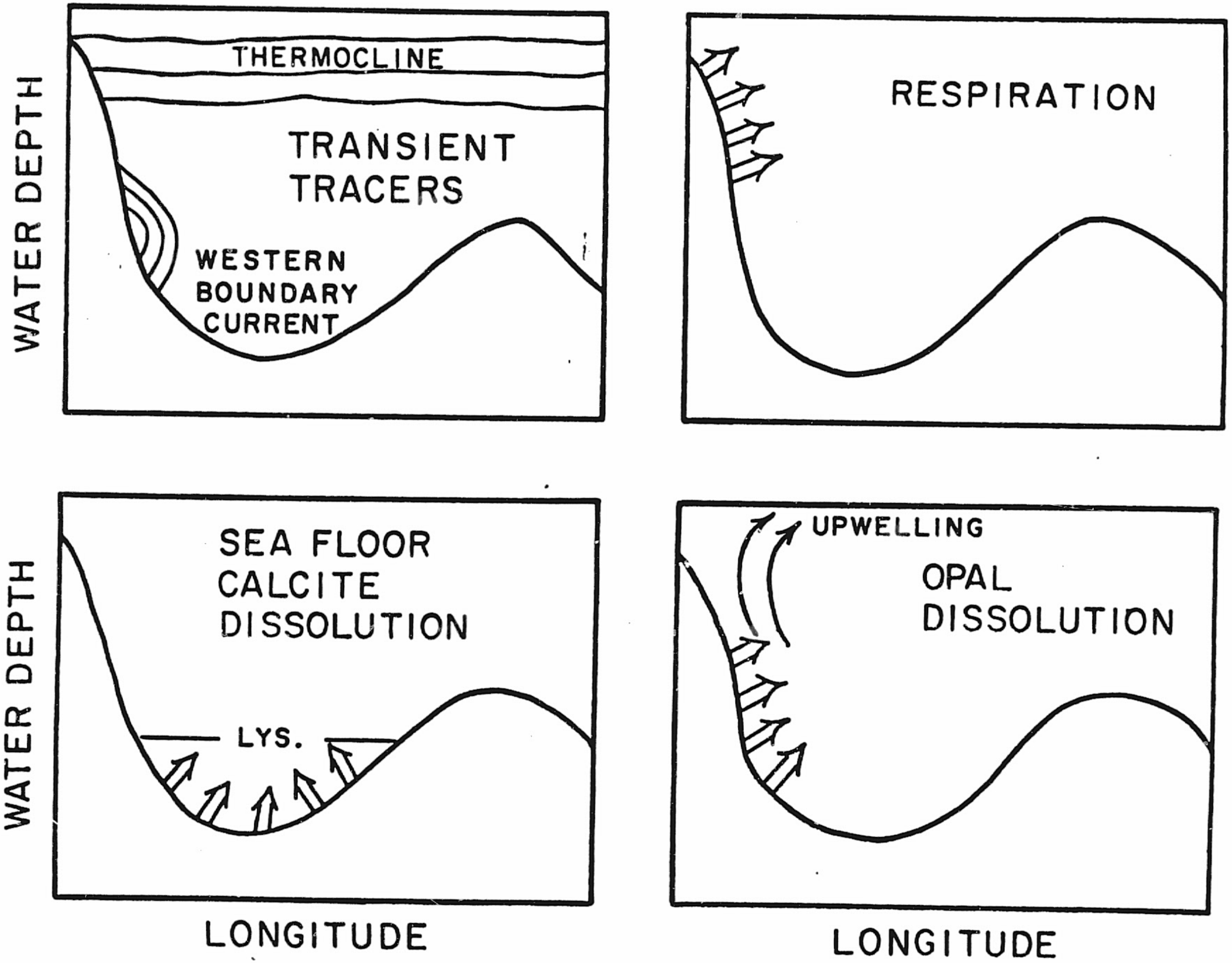

THERMOCLINE
TRANSIENT TRACERS
WESTERN BOUNDARY CURRENT
WATER DEPTH
RESPIRATION
SEA FLOOR CALCITE DISSOLUTION
LYS.
WATER DEPTH
LONGITUDE
UPWELLING
OPAL DISSOLUTION
LONGITUDE

FIG. 20

The sediment trap approach involves measuring the amount of $CaCO_3$, opal, and soft tissue caught over a period of time (usually one year). The problems with the technique are two: 1) organic material is eaten while in the trap and/or organisms invading the trap for food are killed and contribute to the catch; 2) horizontal currents lead to over or undertrapping. Efforts are in progress to minimize both of these difficulties.

The benthic flux chamber is a helmet placed over the sediment-water interface for a fixed period of time (several days). Respiration and dissolution on the interface and within the sediment beneath the helmet raise its NO_3, PO_4, ΣCO_2, H_4SiO_4 alkalinity and its lower O_2 concentration. Ray Weiss at SIO has recently completed the construction and testing of such a device. It is now ready for use in the deep sea. Ken Smith has an operational system for O_2 utilization measurements.

As stated above, by subtracting the accumulation rate of the constituents of interest in sediments beneath a sediment trap from rain rates measured in the trap an estimate can be made of the rate at which each constituent is being released to the water column beneath the trap. If in addition the flux from the bottom is measured using a benthic chamber, then an estimate can be made as to how much of this input occurs at the sediment-water interface and how much occurs in the water column itself. By deploying strings of sediment traps (about 1 per 800 meters) and a benthic chamber at a series of stations along basin wide traverses we should be able to greatly increase our knowledge of the respiration and dissolution functions. To make an order of magnitude increase in our knowledge of these functions, about 300 such stations would have to be occupied (i.e., ~ 300 trap mooring deployments and 300 benthic chamber deployments would have to

be made). Each mooring would have to remain in place for one year. If this work were done over a 10 year period, 30 separate mooring systems would be required. Three Weiss type benthic chamber systems would be needed (about one deployment per month could be made with each system). Also several Ken Smith type systems would be needed for more detailed studies along the margins of the sea.

RECOMMENDATIONS

1) WOCE should sponsor a global flux program designed to produce distribution functions for the rate of respiration, $CaCO_3$ dissolution and opal dissolution.

2) WOCE should encourage the funding of programs like TTO designed to carry out highly precise ΣCO_2, alkalinity, O_2, NO_3, PO_4 and H_4SiO_4 measurements with ocean wide coverage once each 15 years.

3) WOCE should measure O_2 and H_4SiO_4 on its routine θ-S tracks.

4) These measurements cannot be made with the needed accuracy without the aid of a first rate group of analysts such as that created by Arnold Bainbridge and Bob Williams at Scripps. The existence of the existing group is currently threatened by funding problems and political problems. As WOCE will require such a group it should seek to preserve this existing group.

These recommendations are summarized in table 5.

Table 5. Summary of Recommendations Regarding Nutrient Studies to be Sponsored by WOCE.

	DISTRIBUTION in SEA				REGENERATION FLUXES			
	Status	WOCE Density Sections	Special Expeditions	Comments	Status	Sediment Trap & Benthic Chamber	Other	Comments
O_2	well known	yes	yes	improve accuracy	poorly known	yes*	-	-
NO_3	well known	no	yes	improve accuracy	poorly known	yes	sediment accum. rate	meas. N_2 flux from sediments
PO_4	well known	no	yes	improve accuracy	poorly known	yes	sediment accum. rate	-
H_4SiO_4	well known	yes	yes	-	poorly known	yes	sediment accum. rate	sediment map
ΣCO_2	known	no	yes	improve accuracy	poorly known	yes	sediment accum. rate	productivity map
Alk	known	no	yes	improve accuracy	poorly known	yes	sediment accum. rate	-
pCO_2	known	surface water only	yes	-	-	-	-	-

*O_2 requirement to combust sediment trap organics.

CLIMATE MODELS RELATED TO STREAM 3: OUTSTANDING TECHNICAL PROBLEMS

Kirk Bryan

Geophysical Fluid Dynamics Laboratory/NOAA

Princeton University

Princeton, New Jersey 08540

1. Introduction

Models must be able to simulate climate variability on time scales from a few years duration to over a century to be suitable for the objectives of the World Climate Program Stream 3. Special models of the ocean and pack ice are also needed to process the large volumes of data received from new satellite observations and special surface platforms launched as part of WOCE. Several outstanding reviews have recently appeared. Models related to the CO_2/climate question are treated in a U.S. National Academy Report (N.R.C.,1982). A recent workshop has made an excellent report on the present status of sea ice modelling (WCP #26,1982). Ocean models are discussed in N.R.C.(1980) and the Tokyo Study Conference Report (World Climate Program,1982b), and some excellent recommendations are made for future development. Since the broader aspects of modelling related to the Stream 3 objective have been adequately covered in these reports, it seems unnecessary to repeat this material. It is perhaps more appropriate from the standpoint of the JSC/WGNE to go beyond these summaries and examine a few outstanding technical problems of the advanced climate models in more detail. Since various branches of climate modelling require scientists trained in different disciplines, information on numerical methods may not be widely shared. In particular, workers in ocean and pack ice modelling may not be aware of all the numerical methods explored by the much larger community of atmospheric modellers. The goal of this report to stimulate a greater cross fertilization of ideas.

2. Atmospheric Models

The most advanced atmospheric models available for Stream 3 are the atmospheric general circulation models which are not unlike the numerical forecasting models now being used at major forecasting centers. Thus, these models are not really different from those to be used in Streams 1 and 2. The special problems unique to Stream 3 arise in connection with coupling these models to models of the ocean and cryosphere.

3. Ocean Models

It has been recognized since the inception of the WCRP that models of the ocean are presently a weak link in understanding the entire climate system. Many of the field programs of the WCRP are designed to provide the global data sets which will set the stage for a vast improvement of ocean models. Clearly the high priority that the Tokyo Report attaches to ocean models is well justified.

To fix ideas, let us set down the governing equations of an ocean model. Let $\underset{\sim}{u}$ represent the horizontal velocity, and ∇ the horizontal gradient operator. The equations of motion with the Boussinesq assumption, and the continuity equation may be written:

$$\frac{d}{dt}\underset{\sim}{u} + f\underset{\sim}{k} \times \underset{\sim}{u} + \nabla P/\rho_0 = \underset{\sim}{F} \tag{1}$$

$$\rho g + \partial_z P = 0 \tag{2}$$

$$\nabla \cdot \underset{\sim}{u} + \partial_z w = 0 \tag{3}$$

Here f is the Coriolis parameter. ρ is the density and F represents an unspecified closure approximation for smaller scale, unresolved motions. The "so-called" metric terms involving the relative angular velocity about the Earth's axis of rotation of the currents is included in atmospheric models, but can safely be neglected in ocean models.

The equation of state for sea water is a complicated expression involving temperature, pressure and salinity.

$$\rho = G(p, \theta, s) \tag{4}$$

It is convenient to use potential temperature as a predicted variable, rather than temperature itself to allow for the effects of compression. The predictive equations for temperature and salinity may be written,

$$\frac{d}{dt}\begin{pmatrix} \theta \\ s \end{pmatrix} = \begin{pmatrix} Q \\ \sigma \end{pmatrix} \tag{5}$$

Here Q and σ represent the closure approximations representing the effects of mixing by unresolved motions.

Time scales of the thermohaline circulation

The Earth's climate on the time scales of Stream 3 depends on the exchange of heat with the main thermocline of the ocean at all latitudes. Thus the processes which determine the steady state structure of the upper ocean must be taken into account. This is in contrast to ocean models used in connection with Stream 2 objectives which are only concerned with ocean heat content changes in the upper one to two hundred meters of the oceans, principally in the tropics. Models in which the density structure is specified a priori, such as quasi-geostrophic models, are useful for studying the response of the ocean to wind stress patterns, but are not easily adapted to the objectives of Stream 3. We require models which predict the thermocline structure and ocean heat balance. Equilibrium solutions from such models may then be perturbed by changing boundary conditions to study longer time scales of climate variability. The global data sets for geochemical tracers collected by the U.S. GEOSECS program and cooperating expeditions provide a very valuable means of vali-

dating water mass formation in the models.

Let us compare the problem of finding climatic equilibrium in ocean models with that of finding climatic equilibrium in atmospheric models. State-of-the-art atmospheric general circulation models resolve synoptic scale motions, and contain fairly detailed models of radiation and the hydrologic cycle. Such a model will reach climatic equilibrium in 6 to 12 months of integration with respect to time. The heat capacity of the atmosphere is less than 1/1000 that of the ocean, while the heating anomalies driving the two systems to equilibrium are essentially the same. Thus complete equilibration of the ocean, including the deep ocean, will require from 500 to 1000 years. Radioactive tracers pro. de independent confirmation of these time scales.

Is it feasible for ocean models to be numerically integrated to equilibrium in a method directly analogous to that used for atmospheric general circulation models? To gauge the effort required to make a numerical integration over the equivalent of 500 to 1000 years we can examine the wave and current velocities shown in Table 1. A dispersion diagram for the ocean from Hasselmann (1982) is shown in Fig. 1.

Type	Atmosphere	Ocean
Gravity Wave		
External	300	200
1st Mode	100	3
Currents		
Jets	150	1.5
Interior	-	0.2

Table 1: Velocities which may limit the time step of a numerical integration in an atmospheric or ocean model in units of m/s.

Table 1 shows that there is no real difference between the equivalent depth of external modes in the atmosphere and the ocean. If external gravity waves are fully resolved the limitations on time step would be the same in both models. When the external mode is filtered out of the ocean model, the situation is somewhat better. The time step for ocean models can be 30 times longer than that of the atmosphere, but even this advantage does not compensate for the very great difference of 1000 in the natural time scales. How is this difficulty in marching ocean models to equilibrium to be overcome? One approach is to perform variability experiments using solutions which are not in climatic equilibrium with respect to the deep ocean. This is acceptable if

it can be demonstrated that the non-equilibrium "drift" is not large enough to affect the variability experiments. In general, this is difficult to do and may throw in doubt the whole credibility of the model experiment. Clearly there is a very important technical problem to be overcome for the Stream 3 objectives. A great deal of model development remains to be done. Two approaches that have been proposed to make ocean models more efficient are outlined below.

The World Ocean as a mosaic of highly filtered models

Hasselmann (1982) has recently proposed a coupling of highly filtered models for different regions of the ocean. One sector would handle the slow physics of the ocean interior. Another sector would take care of the faster flows in western boundary currents. A special model is also needed for the equatorial region. Figure 1 from Hasselmann's paper shows that gravity waves and external Rossby waves tend to be in a range of frequencies and wave numbers outside of the region of climate interest. To eliminate these waves, all time dependence is filtered out of the external mode, and velocity in the internal mode is constrained to be geostrophic. Only ultra long, nondispersive Rossby waves are possible in this system. The slow physics time step is limited only by the interior velocities of the order of 0.1 or 0.2 m/s. Thus the time step of the interior sector can be of the order of 1000 times longer than in most atmospheric models, enough to roughly compensate for the vast difference in natural time scales.

The method has been partially tested (Maier-Reimer et al. 1982). Starting with the observed temperature and salinity structure as an initial condition,

an ocean model is successfully integrated over a fifty year period with a horizontal resolution of 5°x5° of latitude and longitude and 5 layers in the vertical. There are some distortions in the model thermocline, but these seem to be associated with the first order finite differencing used rather than the essential concept of coupling different regions.

Distorted physics approach

Another method has been used for studies of coupled ocean-atmosphere models (Bryan, Manabe, and Pacanowski, 1975; Bryan and Lewis, 1979). The basic model consists of the primitive equation, with external gravity waves filtered out, and external Rossby waves treated implicitly. Local time derivatives of internal mode velocity components are multiplied by an arbitrary factor, α. If this factor is unity, then the full phsyics for the internal mode is retained. If this factor is made larger than one, the physics of the model is distorted to allow a much more rapid time-marching to equilibrium. Once equilibrium is obtained the parameter can be reset to unity for variability studies.

To illustrate the effect on waves of this procedure, consider a linearized form of the model given by (1)-(5). For a horizontally uniform depth and stratification it is possible to express variables in a set of vertical modes (Moore and Philander, 1977),

$$\underset{\sim}{u}, p = \sum^{N} (\underset{\sim}{U}_n, \rho_0 g h) Z_n \quad . \qquad (6)$$

The governing equations for each mode reduce to a simple set of shallow water equations.

$$\alpha \, \partial_t \underset{\sim}{U}_n + f \underset{\sim}{k} \times \underset{\sim}{U}_n + g \nabla h_n = 0 \tag{7}$$

$$\partial_t h_n + H_n \nabla \cdot \underset{\sim}{U}_n = 0 \tag{8}$$

The undistorted case is equivalent to $\alpha = 1$. At midlatitudes there is a large gap between the frequency of internal gravity waves and the frequency of internal Rossby waves. Thus the beta effect can be neglected in a dispersion relation for internal inertial-gravity waves. From Bryan and Lewis (1979) we have,

$$\underset{\sim}{U}_n, h_n \sim \exp[i(kx + ly - \omega t)] \tag{9}$$

and

$$\omega^2 = (f/\alpha)^2 + (gH_n/\alpha)(k^2 + l^2) \tag{10}$$

Rossby waves can be calculated using a quasi-geostrophic formula,

$$\omega = -k\beta/(\alpha k^2 + \alpha l^2 - f^2/gH_n) \tag{11}$$

β is the gradient of f with respect to latitude. In Fig. 2 the effects of the distorted physics can be seen for both gravity and Rossby waves. Plots are made for an alpha factor of 20. Note that there is a shift toward lower frequency and in the case of Rossby waves towards lower horizontal wave numbers. The distorted physics approach works in the same manner for equatorially trapped modes, slowing down the Kelvin mode wave speed by a factor of $1/(\alpha)^{1/2}$.

If the free waves are slowed down by the distorted physics shown in Fig. 2, the only limitation on the time step is due to advection by ocean currents. In most cases, surface currents are much faster than deep currents. Bryan and Lewis (1979) show how the upper ocean and the lower ocean can be treated separately, and coupled together. In effect, Bryan and Lewis (1979) achieve the same type of economy by separating slow and fast physics in the vertical, as Maier-Reimer et al. (1982) do by separating slow and fast physics in the horizontal plane. In a CO_2/Climate study Bryan et al. (1982) achieve a speed up factor of 25 for the deep water relative to the surface water. This is very similar to the speed up factor achieved by Maier-Reimer et al. (1982). It is possible that still further economies can be achieved by combining elements from both methods.

Pack ice models

The WMO/CAS meeting of experts on sea ice (WCP-26,1982a) noted that existing data sets and new data sets to be collected in the coming year could form the basic of substantial improvement in ice models in the near future. A controversial aspect of pack ice models is the level of detail needed in the ice dynamics. Hibler (1979) has shown that synoptic scale wind events are coupled to the formation of open leads. This effect can cause a substantial augmentation in the growth of ice. On this basis it can be argued that a climate model which resolves individual synoptic events in the atmosphere should also include the corresponding level of detail in a model of pack ice. This view is certainly reasonable, if the main focus of an investigation is climate variability in the polar regions.

Including ice dynamics poses special numerical difficulties. Hibler (1979) gives the equation of motion for pack ice,

$$m \frac{d}{dt} \underset{\sim}{u} + m \hat{k} f \times \hat{u} + m g \nabla h = \tau_A + \tau_W + \partial_j \sigma_{ij} \tag{12}$$

here m is the mass per unit area of the ice, ∇h is the tilt of the ocean surface, and A and w are the stresses exerted by the atmosphere and ocean on the ice. σ_{ij} is the internal stress tensor of the ice, where

$$\sigma_{ij} = \sigma_{ij}\left(\dot{\varepsilon}_{ij}, P\right) \tag{13}$$

The stress tensor is a complicated function of the strain rate, ε_{ij} and the ice strength P, where P in turn depends on the ice thickness. Hibler's (1979) model allows for a small resistance to deformation for diverging ice, and a large resistance for converging ice. The particular details of the stress tensor formulation need not concern us, only the fact that it is highly variable in space in time. In effect a realistic ice dynamics rheology implies something analogous to a viscosity which can range from nearly zero to very high values. Hibler (1979) employs an implicit method for including the internal ice stress, which avoids the necessity for taking short time steps in regions of high effective viscosity. On the other hand, the implicit equations are complex, and finding an efficient way to solve them on a modern

vector computer is not easy. Considerable attention should be devoted to finding improved methods for solving the numerical problems posed by the rheology of sea ice models.

4. Discussion

The variability of atmospheric climate models has been studied by numerically integrating the general circulation models to a steady state, and then perturbing the external boundary conditions. The time-dependent response is then analyzed in detail. The same approach for ocean models must be modified because of the extremely long, natural time scales involved, nearly three order of magnitude greater than those of the atmosphere. Alternative methods for achieving equilibrium solutions have been proposed by Hasselmann (1982) and Bryan and Lewis (1979). Essentially these methods isolate the "fast" and "slow" physics of ocean circulation. In some cases the "fast" physics is filtered out, or artificially slowed down. In other cases regions of "fast" physics such as western boundary currents are numerically integrated separately from interior regions of "slow" physics. During the iterative process the two regions are coupled asynoptically. In the same way the "fast physics" of the upper ocean may be separated from the "slow physics" of the deep ocean. The result may delete or seriously distort natural time-dependent variability. This is not of any concern, however, since the main goal is to accelerate convergence to a balanced equilibrium. Specific methods tested are only a small subset of the many possible approaches which remain to be explored.

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Figure Captions

Fig. 1. Dispersion curves for a horizontally homogeneous ocean. Radius of deformation of the first baroclinic mode is50 km (from Hasselmann, 1982).

Fig. 2. Dispersion curves corresponding to (10) and (11), showing the standard case (solid line) and distorted physics corresponding to $\alpha = 20$ (dashed line).

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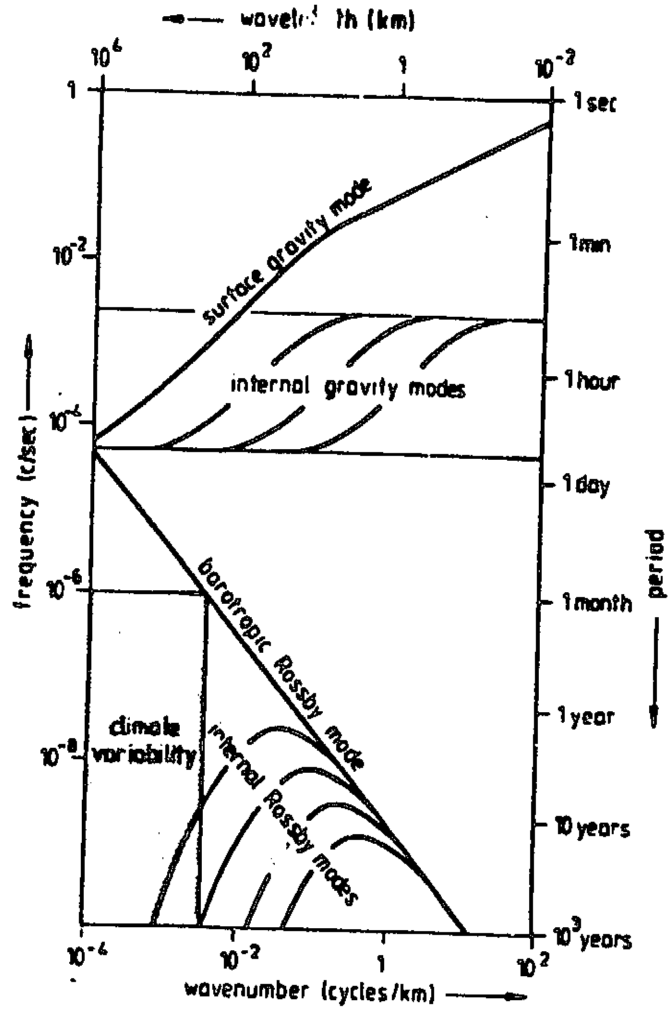
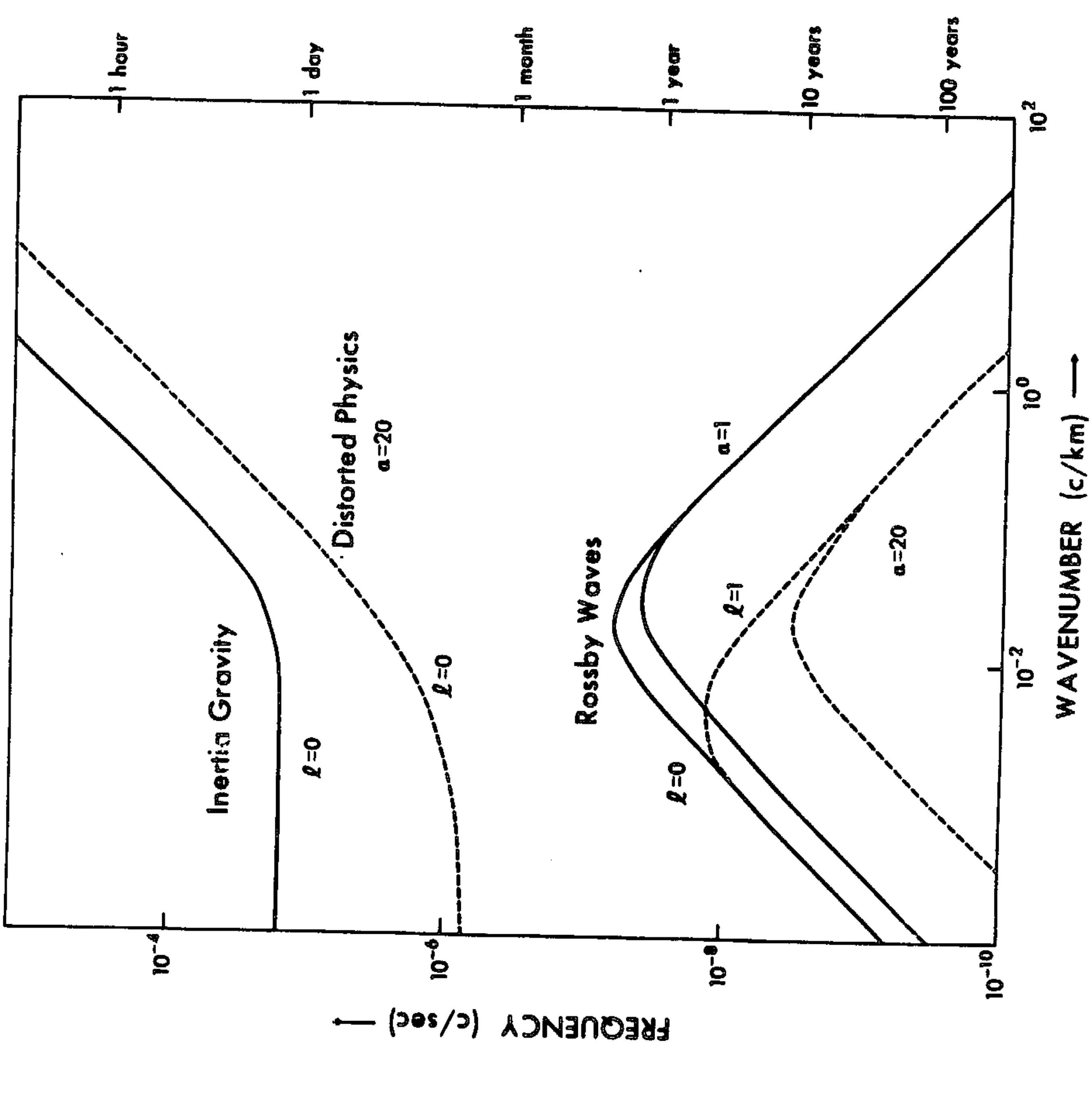

wavelength (km)
10^4
10^2
1
10^-2
1
10^-2
10^-4
10^-6
10^-8
frequency (c/sec)
1 sec
1 min
1 hour
1 day
1 month
1 year
10 years
10^3 years
period
surface gravity mode
internal gravity modes
barotropic Rossby mode
internal Rossby modes
climate variability
10^-4
10^-2
1
10^2
wavenumber (cycles/km)

Fig. 1.


PERIOD
1 hour
1 day
1 month
1 year
10 years
100 years
Inertia Gravity
ℓ=0
Distorted Physics
a=20
ℓ=0
Rossby Waves
ℓ=0
a=1
ℓ=1
a=20
FREQUENCY (c/sec)
10^{-4}
10^{-6}
10^{-8}
10^{-10}
WAVENUMBER (c/km)
10^{-2}
10^{0}
10^{2}

A GENERIC HEAT FLUX EXPERIMENT

From one vantage point the ocean circulation may be thought of as simply the ocean's way of satisfying the constraints imposed by sources and sinks of fresh water, heat and vorticity at the surface. Almost all the measurements that might be envisioned for WOCE will be useful for transport estimates. The question is; how representative will transport estimates based on altimeter data and hydrographic datataken at single points in time be of the long term mean?

The CAGE experiment was originally thought of as a joint ocean-atmosphere field experiment. The heat balance of an entire basin would be measured using several different approaches. Almost implicit in the original concept was the idea that atmospheric measurements would serve as a check on more uncertain oceanographic measurements. When the CAGE Feasibility Panel cast some doubt on this idea, CAGE was deemphasized as a stand alone experiment.

The observational studies of Bryden and Hall (1980) ans Roemmich(1981) suggest field measurements which could she some light on the time variability of the meridional overturning in the Atlantic. Their work also shows that a program monitoring the overturning would effectively monitor the heat transport across latitude circles. The geostrophic component of the overturning could be measured by instruments at the eastern and western ends of a section, plus some measurements along the mid-Atlantic Ridge. Time-intensive measurements for a least one section seems like a rather obvious pilot experiment for WOCE.

K. Bryan NAS/WOCE/8/83

Drifters in the Study of Oceanic General Circulation

A Position Paper by R.E. Davis

1. Purpose

This note is intended to summarize the author's perspective on using current-following drifters to describe basin scale features of the general circulation. It is predicated on the beliefs (1) that observations of density, chemical tracers, and sea-level will not, by themselves, adequately define the large scale flow field, (2) that this flow field will not be adequately predicted from models without direct observation, and (3) that direct observations of lateral dispersive processes (eddy stirring) are required. Thus direct velocity observations are needed and the question considered is which of these needs can be met using surface and subsurface current-following drifters.

Compared to other direct current measurement techniques, important advantages of drifters are: (1) they intrinsically integrate velocity and thus can be efficient descriptors of large-scale low-frequency currents; (2) they can be economical because deployment is simpler than setting deep ocean moorings, recovery is not required, and lifetimes can be long; (3) drifters provide relatively direct measure of lateral dispersive transport processes. Potential disadvantages are: (1) drifters are not perfect current followers; (2) further technical development is required to achieve the potential of low cost; (3) because drifter sampling arrays cannot be prescribed *a priori* and are dependent on the flow being measured, careful analysis of the data is required.

2. Uses of Drifters

In order to discuss the technical advantages and limitations of current-following drifters, it is important to outline the two principal ways drifter observation might be used.

There are not Eulerian and Lagrangian velocity fields but rather pseudo-Eulerian and pseudo-Lagrangian methods of sampling velocity which produce observations of velocity along different x, t paths. For the purpose of measuring the large-scale long-time average velocity field which we call the general circulation, a large number of velocity observations within a particular region, and spanning considerable time, are required to form an accurate space-time average. To do this one could as well sample for long times at several locations or along many pseudo-Lagrangian tracks. The requirements on any such sampling strategy are that the observations be representative, in the sense that their average is the space-time average desired, and that the cost of obtaining enough samples to approximate this average not be prohibitive.

All observations are subject to some error and thus are not strictly representative; arguments are presented below to the point that drifters, particularly subsurface drifters, are subject to relatively small bias errors. Thus Eulerian means can be determined from drifters if enough observations are taken to filter out geophysical variability. The sampling requirements depend on the scales of the mean field, the eddy variability, and the precision sought. Because variability in the ocean is generally larger than the mean, the most stringent sampling requirement is that for measuring the mean rather than the properties of the variability, such as Reynolds stresses or dispersion.

In addition to the space-time average flow, a satisfactory picture of the general circulation includes a description of eddy transport processes. Eddy fluxes, determined from averages of fluctuation products, is one description. A second is obtained from the statistics of dispersing material particles. The latter is more useful because it can be used to deduce the flux of any passive quantity, whereas the flux of one quantity does permit determination of the flux of another. Useful property transport descriptions, such as one- and

two-particle lateral diffusivities, can be accurately determined from drifters even though they do not exactly follow water parcels.

3. Errors

Surface drifters combine a surface buoyant float with a drag producing element set generally at a depth of less than 200 m. This permits easy tracking, as from the ARGOS satellite system. But the surface buoy is acted upon by wind, surface currents, and rectified surface wave motion, all of which act to produce slippage. Wind set is reasonably well understood, but the errors introduced by wave motion acting in concert with current shear between the surface and the drogue level are difficult to determine short of extensive field testing. In conventional designs it is hard to guarantee that instantaneous errors will be less than 5 cm/sec. The error-producing forcing functions (wind, current shear, and waves) are not truly random so bias errors of 1 to 3 cm/sec must be expected. Assuming these forcing functions have time scales of a few days, slippage errors lead to errors of single particle diffusivities of the order $10^6 cm^2 sec^{-1}$-- substantial, but an order of magnitude or more less than observed. Since the forcing functions are coherent spatially, a smaller error of two-particle diffusivities is expected. Slippage errors are insignificant in determining Eulerian eddy statistics.

Subsurface drifters are not subject to any significant instrumental errors. But because such drifters maintain an essentially constant depth and do not follow vertical fluid motion, they sample in an essentially Eulerian manner with respect to vertical motion while being effectively perfect followers of horizontal motion.

With respect to observing Lagrangian averages, such as horizontal particle dispersion, constant depth buoys are imperfect. The spurious diffusivity associated with vertical water motion on any particular scale is approximately $\langle|\eta\partial_z\hat{u}|^2\rangle T_L$, where $\langle\cdot\rangle$ is an average, $\partial_z\hat{u}$ is the vertical shear of horizontal

velocity on that scale, η is the vertical displacement, and T_L is the Lagrangian integral time scale of horizontal velocity components of that scale. The associated contribution to the horizontal diffusivity is $\langle|\vec{u}|^2\rangle T_L$, so that relative errors are small as the square of displacement over vertical scale. While precise estimates are difficult, the diffusivity error associated with internal waves is probably between 1% and 10% of the wave-induced diffusivity, a very small fraction of the total lateral diffusivity. In most regions, meso-scale variability has an even greater aspect ratio, so that spurious dispersion caused by meso-scale vertical velocity is not significant except in regions of strong currents where density surfaces are strongly tilted and eddy energy large. The most serious problem appears to be that spurious internal wave dispersion, with its short coherence length, may mask genuine two-particle dispersion for small separations.

Because internal wave velocity is divergent in the horizontal plane, wave motion produces bias errors in estimating the space-time average velocity from drifters. The average buoy velocity will not equal the Eulerian space-time average if wave motion causes drifters to spend more time in one phase of the wave than in another. Neither will the float follow the true Lagrangian mean velocity because it does not respond to that part of Stokes Drift associated with vertical motion. If a drifter were placed at just the right depth in a single internal wave mode having a stable propagation direction, velocity amplitude 10 cm/sec, vertical scale 100 m, and displacement amplitude 5 m, the largest error would be of the order 2 mm/sec. Since internal wave energy is apparently isotropic and distributed over several vertical modes, one expects bias errors to be significantly less than 1 mm/sec.

Any horizontally divergent flow can cause a bias between drifter velocities and the Eulerian space-time average flow. To the extent that quasi-geostrophy pertains and the horizontal velocity field at scales larger than the internal waves scale is nondivergent, local concentration of drifters

should not be significant. Frankly, I suspect that this may be the most important sampling problem and that when drifter experiments disclose major convergences, they will be "explained" by theory.

<u>Summary</u>. While drifters are subject to errors, the only significant error in the context of low-frequency circulation studies appears to be a bias of the mean velocity of surface drifters; this may be of the order a few cm/sec. Before a significant study can be based on surface drifters, it is necessary to reduce these bias errors, to quantify them, and to insure they will be maintained by increasing drogue lifetimes.

4. Sampling

In the previous section a number of bias errors were examined. A less obvious bias error of drifter observations comes from the interdependence of the sampling array on both deployment location and the velocity field itself. For example, the average velocity obtained from buoys found south of their deployment location will be biased toward southward flow since the likelihood of finding buoys there is lessened in periods of northward flow. The migration of drifters placed into a field with zero Eulerian mean velocity but variable eddy diffusivity is another example of this kind of sampling bias.

In the context of mapping, avoiding this type of bias requires that drifter density be sufficiently uniform that the number of buoys in a region is not dependent on the flow. Strictly, it is necessary that the density (averaged over all periods from which observations are taken) be uniform over a region much larger than the distance a drifter travels in the integral time scale T_L. This does not require massive simultaneous deployments, but means that drifters must be deployed at many sites before bias is eliminated.

The question of how many drifters over how many years are required to describe the general circulation in a region depends on the precision and spatial resolution required and on the magnitude of eddy variability. Eddy

energy, single particle dispersion, and similar quadratic quantities are relatively easy to estimate once the associated mean is known. Assuming approximately normal distributions of eddy velocity and particle displacement, something like 20 independent samples provide precisions of 30%. Particle-pair dispersion may require more samples since pair separation may not be normally distributed. Mean velocity estimates are subject to a sampling error of the order $\delta U = \sigma_u N^{-1/2}$, where σ_u is the standard deviation of eddy velocity and N is the effective number of samples. Since precisions much less than σ_u are generally required, sample sizes greater than 20 will be required.

The effective number of samples depends on the length of the time series of observation and the integral time scale T_L. Price (1983) finds T_L to be near 8 days at mid-depth over a considerable range of eddy energies. Richardson (1983) finds a value near 10 days for surface currents. Thus 6 buoy-years of data provides approximately 100 observations and an uncertainty of 10% of σ_u. Extreme values of σ_u are 45 cm/sec for surface currents in the Gulf Stream and 3 cm/sec for mid-depth, mid-ocean regions. The uncertainty associated with these values from 6 buoy-years of data provides credible estimates of mean velocity in the respective regions.

The number of buoy-years required to describe the general circulation of a region depends on the spatial resolution required. This, like eddy energy, is highly dependent on position within a basin. A resolution of 1000 km might suffice in mid-ocean while 100 km was needed in strong boundary currents. To map a single level of the North Pacific with 500 km resolution requires some 200 estimates. Thus 1200 buoys-years would be needed to achieve $0.1\sigma_u$ precision in each. This could be obtained from 240 drifters with lifetimes of 5 years. A similar global map would require an effort six times as large.

<u>Summary</u>. The most stringent sampling problem is posed by mapping mean velocity. This requires relatively uniform distribution of average buoy density in space over the region to be mapped and a large number of observations.

Something on the order 1000 drifters with lifetimes of 5 years would provide a useful global map of mean velocity at one level.

5. Cost

Even in moderate quantities, the cost of fielding a conventional satellite-tracked surface drifter exceeds $5,000, one-third for the satellite link transmitter. This assumes modest analysis costs and minimal deployment cost through the use of research vessels of opportunity. Present transmitters are capable of providing more positioning accuracy exceeding that needed in large-scale circulation studies. Efforts are underway at NCAR, Goddard Space Flight Center, and perhaps elsewhere, to develop less capable but much less expensive transmitters. Under NASA sponsorship, efforts are underway to develop less costly buoys, with adequate survivability. If these developments are successful, it may be possible to reduce costs by factors of two to even five.

Subsurface SOFAR-tracked floats might be produced in large numbers for $10,000 each and have potential lifetimes of 5 to 10 years. A major cost of using these floats is the maintenance of a listening station network for tracking; three stations, each costing nearly $50,000 per year to maintain, are required within something like 2000 km of each drifter. This makes SOFAR floats most cost-effective when used in spatially dense arrays. Technical developments are underway at WHOI to reduce tracking costs through satellite relay of data and use of drifting listening stations. At URI an "inverse" SOFAR float which listens to moored sound sources is under development. This may significantly reduce both manufacturing and deployment costs since low-frequency sound sources are large and costly in comparison to receivers. Also under development are drifters which repetitively "pop-up" to the surface for satellite location and then return to depth for current-following; this

precludes observation of high frequency phenomena but eliminates the need to maintain costly acoustic tracking arrays.

Summary. Present manufacturing and tracking expenses cause the cost of fielding a drifter to range from $5000 to $20,000 per year, depending on buoy type, density of sampling array, and the useful life achieved. Developments underway promise to reduce this cost by factors of more than two but less than ten.

6. Summary and Suggestions

Meso-scale Variability

From the general circulation perspective, meso-scale variability is important through it effect on larger scales. In addition to understanding the operative mechanisms, quantitative descriptions of the processes and rates of ocean dispersion are needed, if only to make possible analysis of property distributions at depth.

Drifters are well suited to description of lateral dispersion, first order descriptions have been obtained in some places, and additional programs are underway. The quality of observation is adequate but the number and geographical coverage must be dramatically increased. The limitation to this is economic and largely associated with the substantial cost of maintaining tracking networks.

Improvements to tracking range, increased station longevity, data relay, use of drifting listening stations, and development of "inverse" SOFAR floats all have the potential of reducing the cost of subsurface drifter experiments. Similarly, cost reduction of tracking transmitters may make feasible relatively inexpensive short-lived surface drifters suitable for regional studies.

Aside from descriptions of ocean processes and dispersion, meso-scale and regional drifter studies are providing observations of the general circulation. In the future it is important that advantage be taken of this by extend-

ing the tracking duration of such drifters to provide the most efficient sampling of the general circulation.

Surface Circulation

Compared with deeper motion, near surface circulation is closely linked to wind forcing, includes an important ageostrophic component, and is subject to large and variable vertical shear. Surface drifters have demonstrated utility in the qualitative description of regional circulation patterns, in quantifying eddy variability, and documenting interannual changes of the general circulation. By collecting observations from drifters set for essentially regional studies, Richardson has derived a useful picture of North Atlantic surface circulation.

With an increased interest in short-term climate variability, which in the ocean is presumably most evident in near surface flow, drifters will be very useful in describing the structure of interannual change. When global wind observations become available such studies will be more interesting and profitable. Present surface drifter errors probably do not prejudice such studies; the primary technical problem is cost as it affects achievable observation density. But if the data from such studies are to be useful in eventually determining the space-time average flow, accuracy must be significantly improved.

Subsurface Circulation

The description of subsurface general circulation available from direct observation is limited primarily because float and mooring observations are expensive; subsurface drifters have bias errors vastly smaller than typical sampling errors.

Determination of the weak flows and diffusive transport responsible for the observed distribution of oceanic properties is a central objective of oceanography. Important steps have already been made using floats and more

will follow. The pace is largely limited by cost and logistic difficulty. Thus efforts to reduce the overhead for acoustic tracking and to develop pop-up drifters will, if successful, have positive impact. They must be successful before a serious effort can made to determine directly the flow at some reference level over significant regions.

For the time being, primary effort should be placed in cost-reducing development and exploitation of acoustically tracked floats in localized studies.

Satellite Altimetry

The effectiveness of any future satellite altimeter experiment could be significantly enhanced by appropriate drifter observations. The altimeter is unable to determine mean geostrophic currents because the geoid is imperfectly known. While the geoid could be deduced from long term direct current and density observations, significantly fewer observations would be required to infer it from instantaneous altimeter, current and density measurements. The sampling noise to be filtered out would be only that arising from the different spatial averaging of the component systems, not the entire eddy variability as when means are determined.

Would altimeter calibration best be done using drogued surface drifters or a combination of subsurface drifters and hydrographic surveys? This requires comparison of a potentially large number of surface drifter observations with significant bias error to a much smaller probable number of simultaneous density surveys and subsurface drifter observations. The conclusion hinges critically on the bias errors of surface drifters and these are not well enough known. Accordingly, a high priority must be given to reduction of surface drifter bias errors and documentation of accuracies. Any altimeter experiment plan should address the utility and methodology of using direct current observations to improve the altimeters description of mean surface geostrophic flow.

General Circulation Modeling

(A technical paper prepared for the Workshop on Global Observations and Understanding of the Oceans, 8-12 August 1983, National Academy of Sciences, Woods Hole Study Center, Woods Hole, MA)

D. B. Haidvogel and W. R. Holland
National Center for Atmopsheric Research*
P.O. Box 3000
Boulder, CO 80307

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1. Introduction

The following is a brief summary of the current state of global ocean modeling, an assessment of the progress to be expected in the remainder of the decade, and a discussion of the role(s) which such models may play in future global-scale observational experiments (e.g., WOCE). Realizing that we are only now at this Workshop beginning to explore the feasibility and to define the likely objectives of WOCE, we have attempted neither an elaborate nor a definitive statement. Rather, we have sought to introduce what we feel at this point to be the important concepts and modeling issues as related to a world ocean circulation experiment in the late 1980s. This document is intended to serve, in part, as a focus for further discussion at the Workshop.

The questions we seek to introduce and to address below are the following: First, what is the present status of basin- and global-scale oceanic general circulation models, and what can be expected in the immediate future (5-10 years)? Second, what is our present level of understanding of important oceanic processes, and how might this understanding (or lack of it) influence our WOCE-related modeling? Third, what positive interaction may be anticipated between existing and projected models and the WOCE observational program?

In limiting ourselves to an oceanic, global-scale perspective, we have necessarily overlooked a wide variety of models and modeling activities which will be important to future large-scale oceanographic measurement and interpretation. For example, regional models may be

used to examine ocean response and dynamics in physically and/or geographically unique regions (equatorial, southern ocean, mid-ocean). Process-oriented studies will likewise be needed to refine our understanding of important oceanic processes, and their representation and parameterization within large-scale ocean models. Last, interactive atmosphere/ocean models (both coupled general circulation models, and structurally simpler idealizations) will be particularly useful in studies of water mass formation and attendant θ/S properties.

2. Present and projected status of global models

At present, two rather different modeling strategies are in routine use for application to basin- and global-scale ocean circulation problems. Although both are capable of including such physical/computational elements as bottom topography, wind forcing and complicated basin geometry, they nonetheless yield quite distinct pictures of the global circulation.

The first are high-horizontal-resolution (eddy-resolving) models of the mean and mesoscale components of the wind-driven ocean circulation within a single ocean basin. Typically quasi-geostrophic (QG), and having limited vertical resolution (in general ~3 levels, although a few higher vertical resolution simulations have been done), these models have been instrumental in exploring the origin and properties of the mesoscale eddy field, and the role of eddy-related fluxes of momentum and vorticity in determining the structure of the instantaneous and time-mean circulation. However, these eddy-resolving general circulation models (EGCMs) lack the active thermodynamic processes associated

with water mass formation and meridional circulation.

The second class of global ocean models suppresses eddy dynamics through the use of low horizontal resolution, but includes thermodynamic effects and related processes (e.g., deep convection). Generally based on the primitive equations (PE), these oceanic general circulation models (OGCMs) are physically and numerically more complicated than the quasi-geostrophic eddy-resolving models, and require heightened vertical resolution (5-15 levels being typical). However, with a coarser horizontal grid, the models may be extended to cover the world ocean. These coarse-resolution OGCMs have been applied to studies of thermocline formation and maintenance, chemical tracer modeling, and ocean/atmosphere coupling on decadal and climate time scales.

In the future, we believe that the primary advance in global-scale ocean circulation modeling will be associated with the construction and utilization of models which incorporate both active eddy and thermodynamic processes, and thereby simultaneously include the wind-, eddy-, and thermally-driven components of the ocean circulation. Of particular interest will be the richness of behavior that arises in these new models, and the extent to which the resulting simulated flow may be understood by appeal to earlier EGCM and OGCM results. (I.e., are the eddy-induced and thermally-forced circulations additive in any simple sense?)

Secondary improvements in modeling technique can also be envisioned during the next half decade. Efficient new numerical methods (multigrid, nested-grid, composite mesh) are now available which will offer

substantial improvement in efficiency, particularly for models within complicated geometries. Improved representations of, and parameterizations for, processes such as deep convection, subgrid-scale effects, and air/sea coupling would also be desirable. Finally, alternate physical models (e.g., balanced models) will be available, and may in some circumstances be advantageous.

The successful development, implementation and application of an eddy-resolving, thermodynamically active global ocean circulation model over the next 5-10 years depends primarily on three factors: the availability of a sufficient computational resource, the adequacy of manpower levels (both support and scientific), and continued improvement in our understanding and parameterization of certain critical ocean processes. The last of these is discussed in the next section.

With respect to the new modeling being discussed here, the presently available human and computational resources are inadequate. Although the precise level of manpower necessary for the design and application of these models to WOCE-related problems is difficult to quantify, it certainly exceeds the presently available levels. In addition, of the factors noted above, the identification and training of new ocean modelers and dynamicists is potentially the most serious concern, and may more than any other factor dictate the rate of progress in these new modeling areas.

Estimates of the required CPU resource can more easily be made. The table on the following page compares the operational parameters and computational cost of several current/projected models. Although not

Model	Domain (km)	$\Delta x/\Delta t$ (km/hr)	Levels	Integration time (yr)	Required CPU (hrs on CRAY1)	Comments
QG EGCM	North Atlantic	28/4	3	10	5.5	
QG EGCM	box (3600 x 2800)	20/2	8	10	10.5	
PE EGCM	box (4000 x 4000)	20/0.33	3	10	96	Layered model; no active thermodynamics
PE OGCM	world	100/2	5	100	500†	No eddy dynamics
PE EGCM	box (4000 x 4000)	28/0.5	10	100	0(2000)	
PE EGCM	North Atlantic	28/0.5	10	100	0(4000)	Storage becomes a potential problem here
PE EGCM	world	28/0.5	10	100	0(20000)	

Sources: (1) Run-times for Holland models on NCAR CRAY-1 system.

(2) Ocean Models for Climate Research: A Workshop (1980), National Academy Press.

†This figure can be substantially reduced (factor of two or so) by the use of appropriate acceleration techniques to achieve steady-state.

specifically quantified in this table, availability of fast-access memory may also place a significant constraint on the feasibility of using these global-scale models.

The figures quoted in the table are, of course, approximate. Note also that anticipated increases in computer speed (perhaps by a factor of 5 for a class 7 machine) will reduce the number of CPU hours necessary for these sorts of model studies. Future enhancement of computer facilities and staff devoted to ocean circulation modeling is projected at several institutions [e.g., at NCAR, GFDL, NORDA and GLAS; see also An Assessment of Computational Resources Required for Ocean Circulation Modelling (1982), National Academy Press]; if realized, these new resources should allow initial progress on these new model classes. Improved numerical methods may also yield a substantial savings. Nonetheless, the required resource is large. Whereas an eddy-resolving PE simulation of an individual ocean basin should be feasible within the 1980s, an eddy-resolving world ocean calculation will not be possible (even if it were scientifically justifiable).

3. A survey of processes

Our ability to model the ocean circulation in a useful way is intimately connected with our understanding of operative oceanic processes, and our success in including these processes within global ocean models. As a strawman, for subsequent discussion and debate, we propose the following diagram. It attempts to list oceanic processes potentially important to the global ocean circulation, and to indicate which

Lesser Ignorance

Process/Effect	Comments
eddy dynamics	only studied extensively in QG model; what happens in PE EGCM?
equatorial dynamics	
atmospheric fluxes of momentum, heat; ocean-atmosphere feedbacks	what is best boundary value problem for OGCM?
topography	can be studied; but has received insufficient attention
Lagrangian tracers	potentially important for interaction between WOCE and ocean circulation models
ice dynamics	important for WOCE?
convection/water mass formation	present popular parameterization is intuitively crude, but seems to work in a qualitative sense. Is it ok?
surface/bottom boundary layer dynamics	
subgrid-scale effects	Is their precise form important if we have resolved eddy processes

Greater Ignorance

are better understood observationally/theoretically. No claim is made as to which processes are most "important," although this might represent a useful question for disputation at the Workshop. (N.B.: The list is neither fully complete nor unprejudiced.)

All these processes deserve further study. Several are, however, of particular significance to large-scale modeling, and may warrant special attention. In this category, we would include eddy dynamics, ocean-atmosphere feedbacks, topographic effects, and Lagrangian tracer modeling. In addition, we note the important need for further examination of alternate vertical representations, and the appropriate inclusion of isopycnal and diapycnal processes.

4. What can models do for WOCE (and vice versa)

The potential for interaction between the WOCE observational program and the types of models discussed above is substantial. For example:

(a) Experimental design. Basin- and global-scale general circulation models represent a unique means to examine potential observational strategies and data inversion techniques, and to select the most appropriate approaches given program objectives, and organizational and financial constraints. Models have been used in this way for advance planning in programs such as POLYMODE and FGGE. Experimental design studies related to drifters/floats, tomography, altimetry, and satellite-derived winds are already underway at a number of institutions [e.g., Rizzoli (MIT), Riser (U of W), Marshall (Oxford), and others].

(b) Forcing functions. Apart from purely diagnostic models (not discussed here), an ocean circulation model takes the form of an initial/boundary value problem for the current and density fields. In particular, such a global model is driven from above by exchanges of momentum and heat with the atmosphere, or by exchange of mass/momentum/heat across open boundaries (if a single ocean basin is being considered). Of these forcing functions, several are available from existing and projected instrument systems:

Forcing Function	System
surface winds	scatterometer, drifters
SST	satellite, drifters
height field/pressure	altimeter, buoys
open boundaries	
density	XBT, CTD, tomography
currents	current meters, floats and drifters

At present, we do not know with any certainty the amount of data which will be required for these models (i.e., what space/time sampling will be necessary). Of course, this will ultimately depend on overall program objectives. However, model simulations will be essential to deduce the implications of various alternative sampling schemes.

(c) Data for validation. It is of course understood that our models are approximate in many respects. It is important therefore to acquire data sets suitable for model comparison and validation. As often as not, such comparisons show where the models may be weak and in

need of improvement; however, this is as important a result, though hardly as satisfying, as good agreement between data and model. For validation purposes, we suggest the following as being particularly useful: maps of the "mean" circulation (particularly the deep, potentially eddy-related, flow); global maps of eddy statistics (eddy kinetic energy, heat content, height field); maps of thermocline variability; and maps of the distribution of various Lagrangian tracers. The latter, in fact, may offer the most sensitive index of the general circulation, and may be much harder to model correctly than other quantities such as eddy KE.

(d) Assimilation. Models may provide an important framework for examining the composite data set arising from a global-scale observational experiment. Recent studies with oceanic as well as atmospheric circulation models have indicated the feasibility and advantage of producing optimal combinations of observed and simulated fields using appropriate assimilation techniques. Such a model/data composite, if successful, serves essentially to extend the data (i.e., "fill in the gaps"), thereby allowing a more direct and detailed statistical and dynamical interpretation of the data. Given the likely irregular and/or sparse nature of data sampling within any feasible global observational experiment, however, the assimilation problem for the large-scale ocean circulation will be a difficult one. Nonetheless, success in this area is likely crucial to the success of WOCE.

PHYSICAL OCEANOGRAPHIC INSTRUMENTATION DEVELOPMENTS FOR A WORLD OCEAN CIRCULATION EXPERIMENT

By Robert Heinmiller

With contributions and advice from:

Ed Brainard - ENDECO
Al Bradley - WHOI
Neil Brown - NBIS
Bob Chase - WHOI
John Dahlen - Draper Labs
Russ Davis - Scripps
Andy Santos - Horizon Marine
Don Dorson - URI
Bill Large - NCAR
Jim Hannon - Sippican
Jim McWilliams - NCAR
Peter Niiler - Scripps
Tom Rossby - URI
Tom Sanford - APL/UW
Bob Walden - WHOI
Doug Webb - WHOI
Carl Wunsch - MIT

INTRODUCTION

This is a time of great ferment in ocean engineering and instrumentation development. Physical oceanography has always been highly dependent upon, and limited by, the state of the technology. There are several reasons for this. First is the lack of money in the academic sector for engineering development. Second is the lack of good frameworks for such development. Third, the oceanographic instrument market is usually too small to justify private commercial investment in new instrumentation.

There are now, however, in process a number of development efforts, both in the academic labs and in the commercial sector, which hold the promise of a new generation of physical oceanographic instruments by the late 1980's.

This paper assumes a familiarity with the general run of oceanographic instruments available today, most of which have been in use for some years. In a few cases, I will describe possible or planned improvements to existing systems. Rather than giving detailed specifications, I will concentrate on describing in a general way what is already under development, and what is additionally possible within the state of the technology should the scientific community express a need. Lastly, this paper does not deal with remote sensing from satellites, or with tracers, which are both covered by other authors.

The names indicated in parentheses are points of contact, and do not necessarily include everyone associated with a particular developmental effort.

Several general thrusts will become apparent from a reading of this document. They include:

- Telemetry of data via satellites
- "Smarter", more flexible, instruments
- Less reliance on ships and at-sea personnel
- Expendable instruments and systems
- More fully integrated data systems

It should be noted that this compilation is not exhaustive. I have no doubt that new instrument systems are under development at academic institutions and in the commercial sector that are not mentioned here. My goal is to give an idea of the general thrust of oceanographic system development and an outline of what is possible, given sufficient impetus from a potential user community and, of course, funds.

LAGRANGIAN INSTRUMENT SYSTEMS

Various lagrangian instrument systems have been in use for many years, and are useful and productive parts of the physical oceanographic instrumentation repertoire. A new generation of lagrangian instrumentation is under development.

SOFAR Floats (D. Webb, T. Rossby, A. Bradley)

SOFAR floats are a mature technology, in use for some years for large scale mapping experiments. They have a useful range of up to perhaps 2500 km. and a lifetime of years, and in many cases collect and telemeter environmental data such as temperatures as well as "reporting" their position. It should be possible to add other environmental sensors if desired.

The chief limitations on the use of SOFAR floats have been the need to be within acoustic range of the network of shore-based Navy listening stations, and associated potential operational problems having to do with access to the stations and priority. These considerations may be largely eliminated by the use of Autonomous Listening Stations (ALS), and soon by the RELAYS system, described below.

The ALS is a subsurface mooring which receives and records SOFAR float signals, and which is recovered for processing of the data and determination of the float paths. Thus, the data are not available during the course of the experiment, as they are when shore-based stations are used. With the advent of telemetry from subsurface moorings, this limitation could be removed.

RELAYS (R. Chase)

The RELAYS (Real Time Link and Acquisition Yare System) drifting buoy system is under development at WHOI. The system will collect and transmit, initially via Argos, its own position, measurements from sensors on an 1100 meter cable, environmental data from drifting SOFAR floats, and the time of arrival information on the SOFAR float signals. The endurance of the system is expected to be three years. With a usable SOFAR float range of 2500 km., an array of RELAYS buoys should make possible basin scale experiments without regard to the locations of shore-based listening stations.

The development includes the design of an integrated central data logger/processor, which will interface on the one hand with the Argos telemetry system and on the other hand with the subsurface cable. The cable will carry a suite of addressable sensors, and a common data protocol will be used throughout. Currently planned parameters include currents, temperature, pressure, engineering parameters, and the acoustic signals.

It is expected that suitable sensors will become available in the next few years to enable the addition of conductivity. In addition, it should be possible, if desired, to add atmospheric sensors, such as humidity, atmospheric pressure, air temperature, and winds.

Global Circulation Drifter (D. Webb and R. Davis)

The Global Circulation Drifter, or GCD, is presently under development. The concept is a neutrally buoyant device able repeatedly to ascend to the surface, report position and environmental information via satellite, and descend again to equilibrium depth.

This instrument will allow exploration of very large spatial scales and long time scales and is expected to complement conventional techniques such as moored current meters and acoustic floats, which provide excellent time series data, but limited spatial and temporal coverage. Potential coverage is global.

Initial target performance specifications include:

1) 10 years life, 40 ascent/descent cycles
2) Operation to 2000 dB.
3) Buoyancy energy provided by chemical generation of gas (i.e., sea water and metal hydride)
4) On board microprocessor able to record and manipulate environmental observations
5) Air deployment possible
6) Operation via ARGOS satellite system
7) Commercial price target $6K each, quantities of 50

The major inadequacy of the technique is the infrequency of the position reporting due to the limited number of ascent/descents. A position every three months for ten years is an initial performance target. On the other hand, the GCD may provide a very useful first insight into the movement of large and remote water masses.

RAFOS (T. Rossby and D. Dorson)

RAFOS (SOFAR spelled backwards) is moving rapidly to the operational phase with a Gulf Stream experiment in the fall of 1983.

The RAFOS buoys are small, inexpensive packages which are ballasted for operation at a given depth. The price for the first batch will be of the order of $2500 each, but should come down with larger orders. While drifting they periodically record the time of arrival of signals from moored sound sources, along with environmental information, such as temperature and pressure. Design lifetime is one year. At a preset time, the RAFOS package surfaces and is located by Argos, telemetering the accumulated data. Thus, the complete telemetered data is a record of the path of the float since launch with the environmental parameters along that path.

A limitation of RAFOS is the fact that data is available only at the end of the lifetime of the instrument. It seems clear that the RAFOS and GCD technology should be combined. This would combine the continuous path information of RAFOS with the multiple reporting capability of the GCD.

Flux Drifter (W. Large and J. McWilliams)

The Flux buoy is under development within the DRIFTERS development program. This is to be an integrated free drifting buoy system with telemetry of numerous air-sea interactions parameters. The buoy system will be modular. Present plans are to measure air temperature, wind stress, wind direction, humidity, solar radiation, and a water temperature profile. Wind stress may be measured by recording ambient acoustic noise. Precipitation may also be monitored by this means. A doppler current profiler may also be carried.

It should be noted that the lagrangian behavior of the Flux buoy will be secondary to other considerations, such as providing a suitable platform for the air-sea interaction sensors.

Low Cost Drifter (J. Dahlen)

The Low Cost Drifter, or LCD, is to be an inexpensive, simple, yet sophisticated lagrangian drifter for the measurement of transports in the upper 100 to 200 meters, small enough for one-person deployment from a ship-of-opportunity or aircraft. This development is also taking place within the DRIFTERS framework. The design

philosophy is to produce the best calibrated lagrangian drifter possible, with a well-defined transfer function from water behavior to buoy behavior. Emphasis is on calibration and understanding of the buoy behavior, as opposed to attempting to attain the last few percent of true lagrangian behavior.

To this end, additional parameters will probably be limited to sea surface temperature, but no sensor will be added that compromises the lagrangian behavior or calibration. Similarly, the buoy hull will be kept as low in the water as possible to minimize wind effects. By specifying that data and location transmission may take place at random intervals as infrequent as, say once per week, it becomes possible to use an extremely low profile antenna system, with the expectation that it will be awash most of the time.

Power consumption may be kept low by transmitting only at times when a satellite is in a position to receive the signals. The design is aiming at a several year lifetime.

The cost of the LCD, as well as other lagrangian buoys, will be reduced substantially by the use of a new, lower cost satellite transponder, which will have location accuracy which is not as high as the present units, albeit adequate for general circulation studies. The cost per buoy of the LCD may be as low as $2,000, making possible larger scale deployments.

Wave Zone Drifter (R. Davis)

The Wave Zone Drifter is used for lagrangian studies in the very near surface regime. It uses a shallow (one or two meters) drogue which is supported by, but decoupled from, a set of surface floats. The present version uses a high frequency radio locating system, and has been used in coastal areas, but the drifter could be fitted with an Argos transmitter and used in open ocean areas.

HYDROGRAPHIC INSTRUMENT SYSTEMS

Work is underway in a number of places towards a new generation of CTD systems, characterized by "smarter" electronics, internal recording, integrated checkout and data systems, and integrated water sampling. Such systems will have accuracies and sensor resolutions approaching existing CTD's. They will, however, be cheaper to purchase and cheaper to operate. Lower operating costs will result from integrated data systems which will require fewer trained personnel at sea and ashore.

The internal recording will make it possible to operate from ships without a conducting wire winch, thus facilitating ship-of-opportunity cruises and the use of non-oceanographic vessels.

These instrument packages will have a data port on the side, through which data will be downloaded, diagnostics run, new sampling parameters or measurement scheme loaded if desired, and the batteries charged. The shipboard support unit would be a pre-programmed desktop computer which would have capability for pre-processing and display of data.

A closely related development is that of a fast, free-fall profiling vehicle. This vehicle, which is under active development, would carry a package such as a CTD to a predetermined depth, returning to the surface at high speed (up to 15 knots). During the upward phase of the profile, the vehicle would locate the ship acoustically and "home in" on an acoustic source trailed over the side, landing in a pickup device, such as a net, when reaching the surface. Thus, the winch is dispensed with altogether. All that is needed is some lightweight handling gear that may very well be portable. The vehicle would also include integrated water sampling devices.

While the costs of a CTD plus profiling vehicle might be more than the cost of a conventional CTD today, if a profile to 5000 meters can be carried out in an hour or less, the extra cost of the equipment can be recovered in saved shiptime in one cruise at current daily ship costs.

CTD systems like those described above are well within the state of the art technologically. Moreover, all the elements described above are under development. Some additional efforts will be needed for true system integration.

MOORED INSTRUMENT SYSTEMS

Conventional moored sensors (R. Chase)

There are few truly new moored sensors under development. The improvements will be in the moored platforms (with an eventual impact on the sensor configurations). Except in the most extreme regimes, successful mooring deployments of 18 months are now reasonably routine. The battery lifetime of acoustic anchor releases seems to be the limiting factor in mooring duration for now. Many moorings could probably survive much longer than they are left out for, but recovery is dictated by the need for the data and the battery life of the release. However, microprocessor technology plus telemetry has the potential to make great chances in the way we use these moorings.

Telemetry from subsurface moorings is difficult because of the mechanical vulnerability of the link to the surface, and because of the possibility of the surface link introducing noise into the mooring line and moored sensors, thus negating the reason for using subsurface moorings in the first place. However, experience with sensor cables in programs like RELAYS (above) as well as the possibility of "pop-up" transmitting buoys (J. Dahlen) could result in a capability for telemetry from subsurface moorings.

Telemetry even from surface moorings has not been much pursued because of the unreliability and cost of conducting mooring wires up to the surface float. Again, modern cable technology or acoustic links hold the promise of solving this problem.

This in turn may lead to the development of new sensor packages without recording capability, more modular, and designed for use on such moorings.

One can thus envision a mooring system with a lifetime of two years or more, and telemetry of all data. A significant portion of the cost of a mooring goes into

recovery: the acoustic or other anchor release, often some form of backup recovery buoyancy, and recovery shiptime. If we consider eliminating these costs, we may be approaching the concept of the expendable mooring. The sensors on such a mooring would be cheaper, since they would not have to record, and any data pre-processing would take place in a central processor. We must, of course, add on the cost of the telemetry and central logger. But with shiptime and personnel costs climbing faster than electronics costs, the expendable mooring may be economically viable now or soon.

To this must be added the intangible benefits of telemetry: no need to depend on recovery for data return; and the ability to combine high frequency and low frequency experiments.

The expendable mooring concept should be closely examined. One would have to estimate the costs of changeover and startup, of course. There is a large stock of conventional instruments presently in existence. It is unlikely that these could be converted in a cost effective way.

Moored profiling devices

Profiling Current Meter (C. Eriksen and J. Dahlen) - The Profiling Current Meter (PCM) has now moved to an operational phase. The PCM operates for up to a year, cycling between the top of the mooring (which may be within, say, 20 meters of the surface) and up to 250 meters depth.

The PCM measures currents with a set of electromagnetic sensors, and carries a CTD. Positioning the mooring so that the top comes out at a predetermined (very shallow) depth can be difficult. (The development of a "line trimmer" for later adjustment of the mooring line is feasible, but has not been funded.)

Popup (A. Bradley and W. Schmitz) - The Popup profiler is in the final stages of development, and has passed at-sea operational tests. The popup is a bottom-mounted installation that carries up to 50 expendable canisters. The canisters are released at timed intervals and rise towards the surface, emitting acoustic signals. The

bottom installation includes an array of hydrophones which track the canister to the surface (and for a time on the surface), thus producing and recording a complete three dimensional path, and therefore a current profile.

It is probably feasible to add more canister pallets to the basic installation, thus increasing the sampling lifetime of the system. The overall lifetime may be primarily limited by the life of the acoustic release system.

Acoustic Tomography (C. Wunsch, W. Munk, and R. Spindel)

Ocean Acoustic Tomography is a technique for measuring the field of sound-speed fluctuations within a volume of ocean. In regions of tight T-S relations, or where the disturbances are associated mostly with vertical displacements of the isopycnals, the sound-speed perturbation field may be mapped into density perturbations. The technique was demonstrated in a 300 km by 300 km experiment in 1981. Comparison of the results with CTD surveys indicated reasonable agreement. Further tomographic experiments are now in the planning stages.

A tomographic array consists of several moorings, carrying acoustic sources and/or receivers. (The 1981 experiment operated at 224 Hz.) There are a number of geographical paths between the various pairs of sources and receivers. For each of these geographical paths, several acoustic paths will exist, each representing a different coverage of the "vertical slice" along the geographical path, and each having a different time of arrival at the receiver for a given source pulse.

These separate arrivals (up to a dozen or so in the 1981 experiment) can be resolved in the receiver record, and the arrival times for the whole array inverted to produce the sound-speed perturbation field.

The acoustic pulses can be repeated several times a day, producing a near-synoptic "snapshot" of a large volume of ocean. The horizontal and vertical coverage and resolution of the technique is affected by the water mass properties, overall spatial scales, and arrangement and number of sources and receivers. The addition of reciprocal acoustic signalling between pairs of moorings can add more information to an experiment.

EXPENDABLE INSTRUMENTS

XBT (Sippican Ocean Systems)

Improvements to the existing XBT system include the availability of digital recording systems from several manufacturers. The recorders provide a machine readable record of each trace, with higher resolution as well as a more convenient format than the paper charts. The machine readable format facilitates both data processing and possible telemetry of XBT data.

XCP (T. Sanford, Sippican, and Horizon Marine)

An expendable probe for measuring current profiles is available. It uses an electromagnetic probe to sense the components of current relative to the geomagnetic vector. To avoid the effects of the ship's magnetic field, the probe is wired to a free floating buoy assembly and the data telemetered by rf to the ship. Thus, this technique can be used in an AXCP version, for use from aircraft.

XCTD (Sippican)

An XCTD is under development and should be available by late 1984. Depth capability will be up to 2000 meters. Design goal is 0.05 ppt in salinity.

SHIPBOARD SYSTEMS

A number of systems are in use for making measurements from shipboard, such as underaway continuous thermosalinographs, and the Pegasus, White Horse, and EMVP (electromagnetic) profilers. In addiition, many of the other instrument systems described above necessarily involve shipboard equipment as part of the overall system. Generally, these components can be made, like the in situ components, "smarter" and more flexible through the use of microprocessors or microcomputers.

A new technique under development (T. Sanford) involves motional electromagnetic measurements from towed sensors. Surface electromagnetic measurements can be combined with surface current measurements (from correlation sonar, GPS, or LORAN C) to yield the vertically averaged or barotropic flow. This technique might be used from research vessels or from ships-of-opportunity.

Telemetry from shipboard may also become in some cases, a desirable feature, either for full data reporting or for quality control within a large scale multi-ship operation. Similarly, it may be desirable to integrate a number of shipboard sensors and processors, such as navigation, shipborne environmental sensors, and in situ data.

For both telemetry and data integration, it is desirable to adopt shipboard data systems protocols, such as the SAIL (Serial ASCII Instrumentation Loop) protocol.

TELEMETRY

The reader will have noted that many, if not all, of the instrument systems described above specifiy some form of telemetry, usually through satellites, as a integral part of their operation. There are several reasons for this trend.

In a field where many experiments or observational series can run several years, and in which in situ instruments encounter many environmental hazards, it is becoming less and less acceptable that instruments need to be physically retrieved to recover the data. The drastic improvement in mooring recovery rates in the early 1970's resulted in a drastic increase in the length of experiments. Thus, although ultimately the scientist got more data back, he had to wait longer and longer to see it, and the price paid for a mooring failure was higher.

The increase in the price of shiptime and the associated technical personnel costs has also caused many investigators to consider the possibility of designing systems which are not intended to be retrieved, thus requiring telemetry.

Lastly, the advent of large scale coordinated programs involving many ships, institutions, and countries has suggested a need in some cases for telemetry for coordination, planning and data quality control.

Satellite telemetry and position location is now relatively routine, whether from moored stations, drifting buoys, or ships. Systems are available off-the-shelf commercially. The limitations fall into two categories: money and available telemetry channels. Money limitations, are, of course, always with us.

This is not the place for a detailed discussion of the available telemetry channels or the future needs. I would note, however, that a simple adding up of proposed telemetry activities in the latter half of the decade suggests that there will not be sufficient channel capacity to accommodate the needs of the community, based on what is presently available and planned. (Indeed, by 1990 we could have less capacity available to us than we have now, due to satellite failures.) On the one hand is a possible shortfall in the overall capacity, on the other are limitations or constraints in the individual channels.

There is approximately an eight year lead time in planning and deploying a satellite. Thus, planning should be underway now for the systems for the early 1990's. This is taking place. However, in the context of programs such as WOCE, it appears that the community's needs could easily outrun the plans. Lack of dedicated channels could force us to move to commercial systems, which involves another set of engineering and funding problems.

In the context of WOCE, an early estimate should be made of the range of overall telemetry rates which might be desirable, from the "absolutely necessary" to the "be-nice-if". I realize that this is difficult while the scientific program itself is still fluid. However, the effort should go ahead as soon as feasible. This will necessarily involve estimating other needs outside WOCE for a first order determination of whether the capacity will be there.

OPERATIONAL CONSIDERATIONS

The trend towards expendability will greatly decrease the amount of ship time necessary for some types of operations. (Or, of course, given an fixed amount of shiptime, greatly increase the amount of data that can be taken.) The cost of expendables must be balanced against the reduced ship time, reduced personnel costs, reduced data reduction costs, or some combination of these factors.

The increased availability of "smarter" instrument systems will reduce the number of highly trained personnel at sea, and make ship-of-opportunity operations easier. This means reduced costs for existing oceanographic data-collecting operations, and increased opportunities for operations in countries with less money and fewer trained technical personnel.

"Smarter" packages may also make possible more flexible sampling rates. The sampling rate might be adjusted automatically to take into account features or events in the data being collected. In situ processing can reduce the amount of data storage or telemetry and reduce processing costs ashore. In general, many things that have formally been done in hardware can now be done more flexibly and more cheaply in software.

Telemetry can change oceanographic data-collecting in two ways. First, the scientist is no longer dependent upon the recovery of the instruments to get his data. Gear might be left at sea until it fails, with the data set being accumulated continuously. Secondly, the data storage capabilities of existing instruments are the limiting factor in deployment durations. In many cases, the sampling rates of the instruments are set for the lowest rate to get the desired duration. If data is being telemetered, the rates may be set at the maximum that the sensor response will justify and the telemetry channel allow. Thus, higher frequency experiments can "piggy-back" on lower frequency experiments.

Integrated data systems will reduce the need for trained, specialized personnel for both operations and maintenance. This, in turn will allow the use of such systems by smaller groups and scientists who do not have a sufficient pool of such personnel

to tap. Many instrument systems will be sent back to the factory for routine maintenance between cruises. Having functions in software that are presently in hardware will make checkout and maintenance easier and quicker.

Plans are underway for increased capability for air deployment for lagrangian systems. In some cases, aircraft may be more cost effective than ships. This may be particularly true in some oceanic regions.

In large scale integrated programs, particularly where telemetry is used for data reporting or quality control, data networks and handling systems may need to be implemented. Such systems, based on existing network and computer technology could involve data transmission, pre-processing, preliminary analysis, indexing, and archiving.

CONSIDERATIONS FOR WOCE

There is at present no major thrust to develop instruments to make totally new kinds of measurements. Rather, the trend is to exploit new forms of technology and new ideas to make the traditional measurements more cheaply, more quickly, and more reliably. This will make it possible for oceanographers with less resources in money, specialized ships, and trained personnel to participate in observational programs. In addition, trends toward aircraft and ship-of-opportunity operations may also make some types of work in some regions more cost effective.

Many of these developments are taking place already in various academic institutions and in the commercial sector. To be available routinely for WOCE operations, however, many of the instrument systems will need to be available commercially, in quantity, well-documented, well-supported, and at a reasonable price.

There is a much greater awareness now than in the past of the need to move instruments from the academic development sector to the commercial production sector. This will be crucial if WOCE needs large quantities of a given system. Most of the instruments described above will be available commercially. Large systems such as moorings are an exception (although, even here, it is possible to

get scientific moorings set commercially). It will always be desirable for a scientist to have considerable control over his own observational program. There will be a need for operational personnel, both ashore and at sea, working directly for the interested scientist. However, in many areas it makes sense for the production of instruments, and perhaps some functions such as routine maintenance, to be in the commercial sector.

Some input and participation by WOCE scientists may be needed for some of the developments described above to bear fruit. System integration will be necessary before some instrument systems are truly operational. Specifications can and should be influenced by the needs of the eventual user community. Development schedules might be influenced to meet the time constraints of WOCE. Some operational needs (specialized software, sampling rates, etc.) may not be met unless the engineers involved receive input from users. This input and participation may, in some cases, include the procuring of funds for the actual work.

The scientific plans for WOCE should be framed in terms of what is possible and under development, not alone in terms of what we can do today and have been doing. Once a program is laid out, a subgroup of WOCE should list the resultant needs in instrumentation systems and data systems, identify areas where user input is needed, and outline a formal framework for participation and input.

Other early initiatives might include, as outlined above, an investigation of the possible needs for telemetry and data handling.

WOCE may be of sufficiently large scale to have the "clout" (with both funding agencies and the commercial sector) to influence the future of oceanographic instrumentation development.

AIR-SEA TRANSFER PROCESSES: The Surface Observations

Prepared for the "Workshop on Global Observation and Understanding of the General Circulation of the Oceans"
Woods Hole, MA, August, 1983

W. G. Large
National Center for Atmospheric Research
Boulder, Colorado 80307

1. INTRODUCTION

The processes of air-sea transfer that are relevant to the general circulation of the oceans are sufficiently well understood. There is a circulation driven by the surface wind stress and a thermohaline circulation governed by the surface heat and mass (salt budget) fluxes. The difficult problem then, is to observe these processes on general circulation scales (GCS), that is globally on a monthly or seasonal time scale with a resolution of some hundreds of kilometers. Stress is usually referred to a level in the atmosphere and on GCS, the assumption that it is entirely transmitted through the surface wave field to the upper ocean is likely valid. A complication may arise in high winds _if_ there is significant evaporation of spray in the atmosphere because a parameterization of the latent heat flux from the ocean only accounts for mass lost through evaporation at the ocean surface.

The possibility of satellite observations makes it tempting to reduce direct surface observations to those things satellites cannot measure, but for two considerations. First, higher quality surface measurements will result in better satellite retrieval algorithms. Second, satellites can have their launch postponed and once in orbit there are failures (SEASAT, GOES-West). Therefore, a program of global surface observations, of which stress is the most important, is needed to complement the satellites. A successful program may be able to rely on present techniques in some areas, but will need to advance the state of the art in others. This will be a large and difficult undertaking and care is needed to make the most effective use of the available resources, both financial and human. Can it be done?

2. SURFACE WIND STRESS

The present state of knowledge of the surface wind stress or momentum flux, notwithstanding satellite remote sensing, can be briefly summarized as follows:

a) It could be measured using the dissipation technique, almost routinely from ships or large buoys on the deep sea with about 20% uncertainty hourly. Over GCS the uncertainty would be less, but still more than 10%, because of the assumptions made and constants used.

b) Over GCS, bulk aerodynamic estimates can be as good as 15% provided that the neutral 10 meter drag coefficient is known to within 10%, by formulating it to lie within the solid lines of Figure 1A, and it is corrected for height and stability, and the mean wind speed is as accurate as possible, ±3%. In Figure 1A, vertical bars span the range of observed hourly values, and much of this scatter appears to be associated with non-equilibrium wind-wave conditions. However, this is a short time and small space scale phenomenon, which is mostly accounted for by using a formulation based on averages of drag coefficients measured over a large range of conditions. Despite the averaging, there remains a 10% error due to errors in the data on which any such formulation is based.

c) It seems possible that wind stress or speed could be estimated from the ambient oceanic acoustic noise at frequencies above about 5 kHz, as measured remotely by a hydrophone at a depth between 100 and thousands of meters [1].

d) GCS time averages of the bulk vector wind stresses from ocean weather station (OWS) three hourly observations have compared favorably with refined techniques employing heavily averaged data. Monthly vector averaged winds and an empirically determined transformation that attempts to preserve the climatological means and variances, and that is valid throughout the westerlies, produced monthly stress components whose rms differences ranged from 0.13 dPa at OWS 'M' to 0.22 dPa at OWS 'D', eastwards, and from 0.07 dPa at OWS 'E' to 0.20 dPa at OWS 'P', northwards [2] (1 decipascal = 1 dyne/cm^2). The monthly mean bulk stress vector calculated using the vector averaged wind and its variance agrees 90% of the time to within 15° in direction and ±0.5 dPa [3]. By considering the skewness, four-monthly averaged stress estimates agree to within 0.25 dPa 95% of the time. An extension to larger areas by estimating the variance from data in the U. S. Navy Climatic Atlas and the mean wind from monthly averaged surface pressure maps is in progress, as is an attempt to relate the large scale stress field to variations in Sverdrup transport and sea level [3].

e) The uncertainty in aircraft measurements is at least as large as in good bulk estimates, over large scales, and is too large for scatterometer calibrations [4].

f) Wind speed measurements from moored buoys have improved since JASIN and may be approaching the desired ±3% accuracy [5].

Given the above, some comments relevant to a global observational program are warranted. It is unlikely that the small gain in accuracy

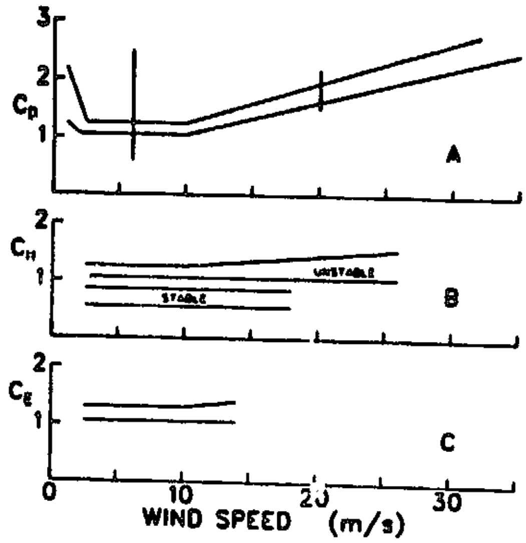

Fig. 1. Neutral, 10 meter, bulk transfer coefficients (x1000) for momentum, C_D, for sensible heat, C_H stable and unstable, and for latent heat C_E. The 'best' GCS formulations as functions of the 10 meter wind speed, ought to fall somewhere between the solid lines.

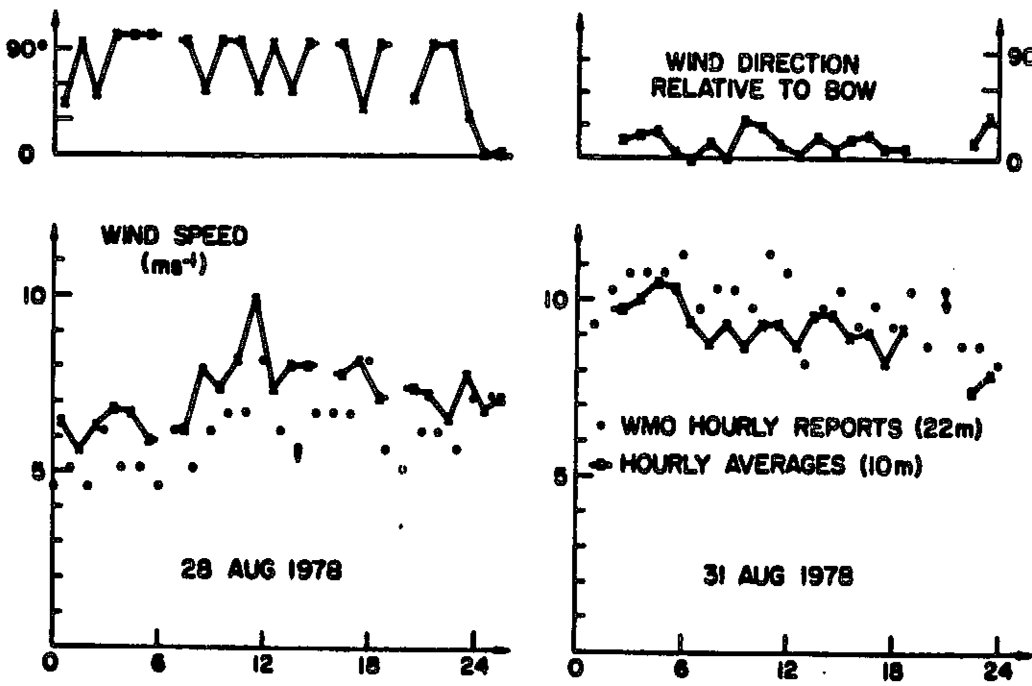

Fig. 2. Wind speed and direction from F.S. Meteor during JASIN, 1978: First, as reported to WMO from a cup anemometer at 22 meters height. The vertical arrows show the speed reduction in moving down to 10 meters. Second, as measured by propeller anemometers with better exposures, shifted to 10 meters (5).

of dissipation measurements over bulk estimates would be worth the extra effort required. Instead dissipation measurements should be used to empirically calibrate other techniques (ambient noise, scatterometer) directly in terms of stress (rather than wind speed), and as a periodic spot check of the observed stress. Even a large concerted effort is unlikely to improve knowledge of the drag coefficient significantly. A more fruitful pursuit would be to improve the wind speed measurements. Figure 2 shows two days of wind measurements from F.S. Meteor during JASIN, 1978. The accuracy of the hourly averages from propeller anemometers are probably within the desired ±3% [5]. It is evident that the standard WMO winds are as much as 2 to 3 m/s or 30% low whenever the wind blows from starboard, and these reports from a research ship are likely better than most. An effort should be made, at least with some ships, to correct or reject, wind speeds according to the relative wind direction. The calibration and exposure of anemometers on these ships could also be improved in many cases. An effort is underway to model the flow around navy and possibly research ships, and to determine wind speed corrections [6].

Figure 3 illustrates that a ship based observational program is well suited over much of the northern oceans [7]. Elsewhere there are few ships and strategically placed moored wind and pressure records and drifting buoy observations seem to be needed. FGGE demonstrated that a one year record of sea level pressure, SLP, and sea temperature is possible from a drifter, but considerable effort is required before accurate wind stresses are feasible. An ambient noise estimation of the stress magnitude is presently being pursued [8] and should be

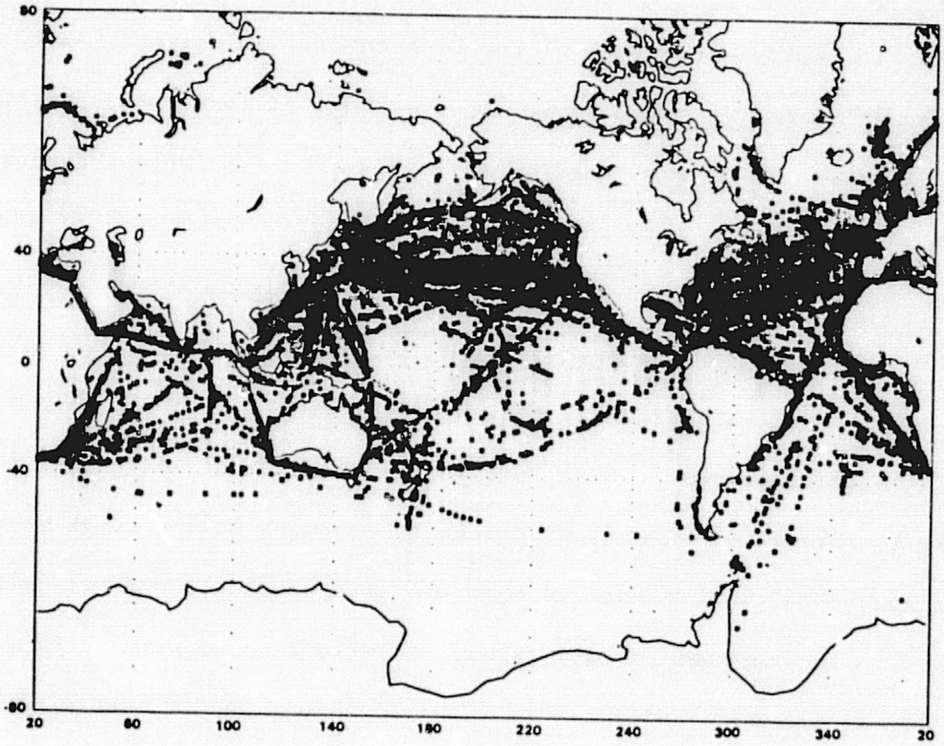

Fig. 3. Distribution of edited FNOC ship reports for first part of November 1979 (7).

commercially available within five years. Can adequate wind directions be derived from SLP maps drawn from buoy and island pressures? If so, the difficult problem of measuring direction with vanes or from the directional noise spectrum, both requiring expensive (in terms of cost and/or power) compasses, is less urgent.

Present efforts are striving to bring the cost of a FGGE type drifter down to around $2500. It is then conceivable that the approximate 200 drifters required to produce accurate SLP analyses (FGGE specifications), could be maintained in the southern ocean westerlies. TOGA will likely provide some of the South Pacific fleet and several good island sites could be utilized. In order to learn how to calculate GCS wind stress from these pressure data, good surface wind records and perhaps satellite winds are needed. Should this calibration procedure yield results of sufficient accuracy (the data to do some preliminary analyses exits), then the regularly sampled and naturally filtered product would be a natural complement to the satellite winds. Pressure data have the added attractions that a lot of processing techniques and expertise already exist and they are of great meteorological interest. Given the low cost of these drifters, it is conceivable that such a program could be implemented in all ship sparse regions, or even globally with few thousand drifters plus ship reports. It could be of great benefit to a much more costly satellite program, which will have its own processing and calibration problems.

3. THERMOHALINE PROCESSES

The balance between the turbulent heat fluxes (latent and sensible) and the radiative heat fluxes (longwave and shortwave) at OWS 'BRAVO' (56°N, 50°W) is shown in Figure 4, along with the total monthly heat flux and its anomaly [9]. The dominant terms change from season to season and year to year with even the sensible heat the largest some winters. The balarce also varies greatly with geographical location [10]. The latent and shortwave heat fluxes are usually the largest terms (in winter at 40°N the longwave flux is of comparable magnitude), but they are of opposite sign and hence the other terms can contribute significantly to variability in the total surface flux. Even within the GATE B/C scale, one-half of the $60W/m^2$ ship to ship variability in net solar flux, as calculated from radiative transfer models, was due to long wave variations [11]. Therefore, all terms need to be considered in a global observational program. In addition, the salinity budget needs precipitation measurements. These requirements are much more severe than those for stress, and it is doubtful if they can be met, globally, without an extensive satellite program.

At present we know the following:

a) All the fluxes, including precipitation, can be measured directly from ships, but only with considerable care.

b) The bulk transfer coefficient for sensible heat, Figure 1B, is different in unstable and stable conditions and the uncertainty in the former is only about 10% at low wind speeds. There is greater uncertainty at higher winds and in stable conditions.

c) The bulk transfer coefficient for latent heat is only known to within 15% at speeds below about 15 m/s, Figure 1C. There have not been any measurements at higher speeds, but a similarity theory extrapolation (yielding a $C_D^{1/2}$ scaling) is probably within 25% over the wind speeds of most flux. If evaporation of spray becomes significant, this C_E will not account for this mass flux.

d) Shortwave radiation estimates based on geostationary satellite images are probably as good or better than those from ships or buoys [12]. The coverage is from 50°N to 50°S and will soon be over all longitudes.

e) Budyko's shortwave parameterization scheme seriously underestimates a model based on OWS 'PAPA' (45°N, 145°W) data both at PAPA and BRAVO [9]. A comparison with other schemes, and an extension to other areas is underway.

f) There are many parameterizations for the longwave radiation, but each is probably only valid in specific areas or seasons [13].

g) Attempts are being made to estimate sea surface temperature [14], total heat flux [15], cloudiness (FIRE program), and latent heat flux [16] from satellite data over the ocean. In each case an evaluation of the success is hampered by the lack of quality surface observations.

Many of the points raised concerning wind stress observations apply to the heat fluxes too, but some additional comments seem appropriate. Again, a great deal is to be gained from improving ship measurements.

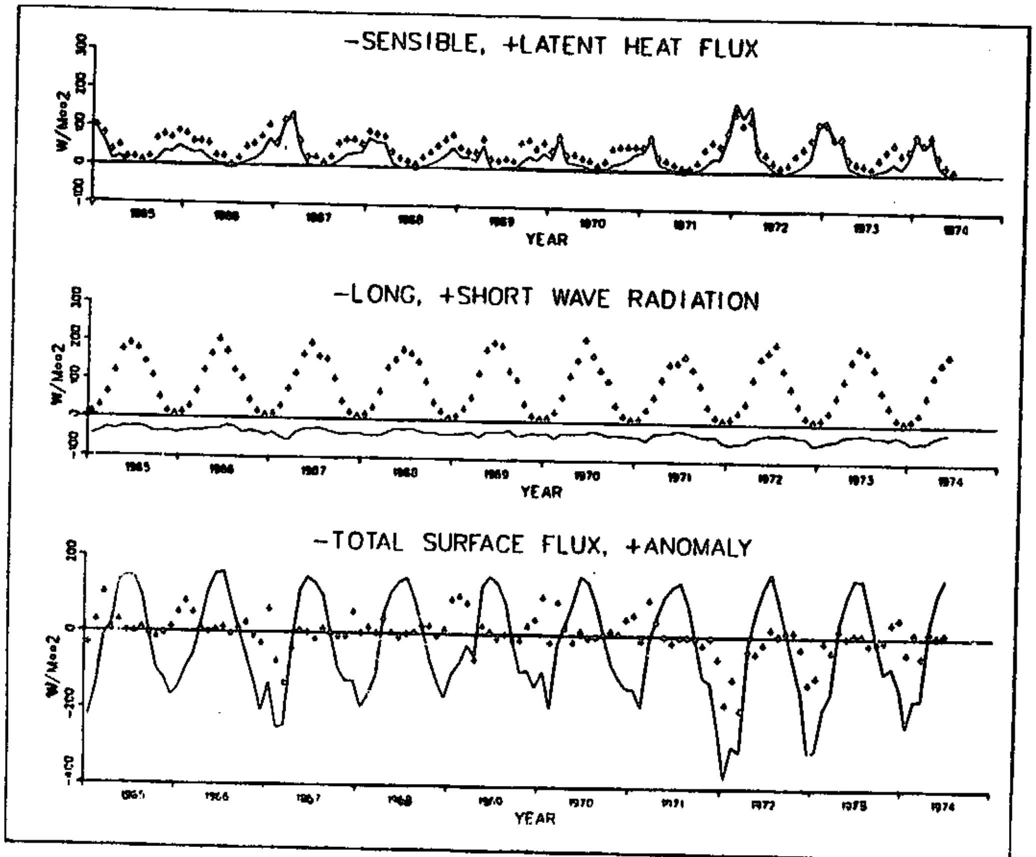

Fig. 4. Surface heat fluxes at OWS 'BRAVO' (56^{o}N, 50^{o}W) 1965-1974, (9).

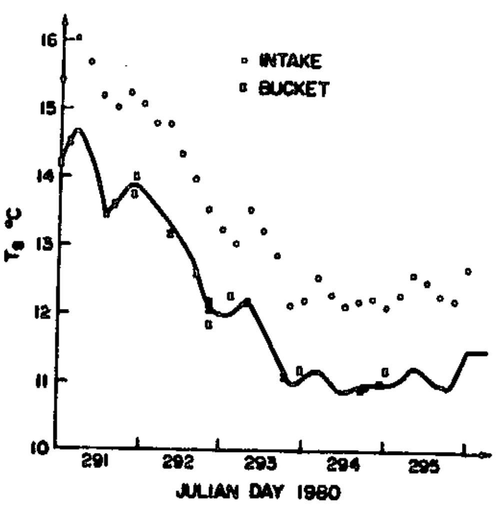

Fig. 5. Sea surface temperature time series constructed from bucket observations and continuous intake temperatures from well inside CSS Parizeau, which are biased about 1.2°C high.

To illustrate by extreme example, Figure 5 displays a large bias (>1°C) in intake sea temperature (measured far inside a ship) versus occasional bucket temperatures. However, with the two data sets it is possible to construct a continuous sea temperature time series, with much better accuracy than the intake temperature and much better coverage than the bucket. A larger scale comparison of intake and BT sea surface temperature [17] shows a 0.4°C bias. Methods of improving air temperature and humidity, at least from some ships, should be investigated.

The observational burden would be greatly eased if monthly (or longer) averages of sea temperature, air temperature and humidity could be used in bulk calculations of GCS turbulent heat fluxes. Again, such refined techniques have been compared to averages of the three hourly fluxes available from the ocean weather ships. Using monthly averages, also of wind speed, the relative error is about 10% in both the latent and sensible heat fluxes [18]. Using monthly averages and the vector average wind, with an empirical wind speed dependent transformation (as in Section 2), the errors in the climatological means are only $3.5 W/m^2$ in sensible and $17.7 W/m^2$ in latent. However, there are large ranges in the rms differences in both the sensible (from $2 W/m^2$ at OWS 'P' to $8 W/m^2$ at OWS 'D') and latent (from $7 W/m^2$ at OWS 'P' to $32 W/m^2$ at OWS 'N') heat fluxes [2]. Geographical areas where such techniques could be applied need to be determined.

In areas with few ships, drifting buoys are capable of giving sea surface temperature (±.05°C) and perhaps salinity and air temperature (~±.2°C) in the near future, but reliable humidity measurements (±3% relative) are contingent upon the success of current developments of

new sensing techniques [19]. Once accurate CGS observations of bulk parameters become feasible, there will be something to be gained by more firmly establishing C_E at high wind speeds and C_H in stable stratification.

Precipitation is an important but neglected process. Satellite estimates inferred from cloud top temperatures are too indirect to be without ground truth and rainfall frequency based on total liquid water content will likely be more valuable. Precipitation produces most of the oceanic ambient noise above about 50 kHz and rates may be quantitatively derivable using a passive hydrophone [20].

Even with the inhomogeneous upper ocean of the JASIN experiment changes in the oceanic heat content over a 100 km x 150 km area surveyed over seven day periods, balanced the surface heat flux (50 W/m^2) to within about the 10 W/m^2 of horizontally advected heat [21]. An advection of 20-30% of the local change in heat storage has also been observed near OWS PAPA (50°N, 145°W) during summer, MILE [22] and fall STREX [23]. It may, therefore, be possible to infer the surface heat flux over certain large ocean areas from the change in heat content with perhaps a correction for vertical advection derived from temperature profiles (MILE) or the curl of the wind stress. Without the dense sampling of JASIN, or an essentially Ekman vertical velocity, or the validity of the MILE assumptions (that the mixed layer isotherms are material surfaces), large errors could result. The surface heat flux also needs to be a large portion of the heat content changes, unlike STREX where it was only about 10%. Nevertheless, the possibility of adapting such an inferred technique to at least limited areas and

perhaps seasons should be explored, because it would seem to be a means of estimating the surface heat flux without recourse to satellite data and it is well suited to drifting buoys.

4. A SUMMARY OF SALIENT POINTS

a) The coverage provided by presently reported ship meteorological observations must be weighed against the desire for more accurate ship measurements.

b) Refined techniques that reduce the observational requirements need to be developed, tested and applied wherever and whenever possible.

c) Drifting buoys are required for remote ocean regions and SLP, Ts and heat content measurements are well in hand. Reliable wind stress and quantitative precipitation derived from ambient acoustic noise would be a major advance.

d) Most of the global shortwave radiation can be estimated from GOES imagery and the possibility of a longwave algorithm is being explored. Image processing is expensive and must be minimized. Satellite cloud climatologies may soon be available to aid the radiation observations.

e) Surface observations are required to complement satellites, and a program could be designed to withstand losing satellite winds for a period of time; however, satellite components of the surface heat and mass flux observations are essential.

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WOCE PLANNING DOCUMENT

PURPOSEFUL TRACER EXPERIMENTS

James R. Ledwell
Goddard Institute for Space Studies
(NRC-(GISS) Research Associate)

Prepared for WOCE Workshop in Woods Hole

August 1983

INTRODUCTION

The ability to tag large regions of the ocean (i.e. $\sim 10^{22}$ cm^3) with artificial tracers lies well within our grasp. A few tons of inert halogenated compounds can be diffused through permeable tubing into a target initial water parcel in a few weeks. The dispersal of the tracer could be monitored using shipboard or possibly even *in situ* electron capture gas chromatography. The lower limit of detectability is of order 3×10^6 molecules, e.g. 3×10^4 molecules/cm^3 in 100 ml of water.

Such tracer experiments could be performed to study large scale current systems, basin wide eddy diffusion along isopycnal surfaces, and, most important, cross-isopycnal mean motions and diffusion inaccessible by other means.

This document presents a brief history of the purposeful tracer idea, the present state of development, plans for the next few years and a sketch of some example experiments of a type that would enhance the WOCE program.

HISTORY

Under the sponsorship of the DOE-CO_2 program in 1978-79 W. Broecker and J. Shepherd undertook a study of the feasibility of purposeful tracers. They concluded that experiments could be done using ^3He as the tracer, since 10 moles of ^3He could be tracked over a large region of the ocean. The reaction of a wide range of physical oceanographers and marine isotope chemists was solicited. The consensus of the community was favorable but three problems emerged:

1) those studying the distribution of ^3He from mantle outgassing and from the *in situ* decay of tritium objected to the superposition of yet another type of ^3He;

2) some of the physical oceanographers feared that the injection would be rapidly broken into a maze of wisps and would thereby defy mapping;

3) a number of people felt that it was imprudent to perform a large scale experiment right away and suggested that the next logical step was to extend T. Ewart's several day thermocline dye patch studies to one month. Fluoroscene dye would be mapped by the free vehicle SPURV as was done in Ewart's earlier experiments.

These suggestions were taken seriously. ^3He was abandoned in favor of methane-21 or some perfluorocarbon. A decision was made to propose one or more one month dye experiments. A theoretical panel was established to ponder the patchiness problem and to consider candidate localities for the first experiments. This group met at Lake Wilderness, Washington in September 1980, and again in Cambridge, England in November 1981. These discussions led many of the interested parties to the conclusion that the one month dye experiments would likely be too short in duration to allow the important questions concerning one to 10 year injections to be answered. As SPURV lacked the range to map a big enough patch to allow useful extrapolation to the scale of the larger experiments it was decided that the one month and several year experiments should be developed separately. Where possibilities for sharing facilities and making joint injections existed the two groups would work together. However, the proposals would be separate and the fate of one would not hinge on the fate of the other.

Broecker and Shepherd made one basic change in the strategy for the long term tracer experiments as a result of a rather strong suspicion by the theoreticians that once the tracer patch had spread to the 30 to 300

km size it would be torn asunder by mesoscale eddy activity and remain long after as a series of wisps in tracer free water. It was decided to hold off any search until several years after the injection. On this time scale the wisps might coalesce sufficiently that valid mapping could be done. This original search would be carried out on ships of opportunity. Only when the location and patchiness of the tracer had been assessed would a dedicated mapping program be launched. Of course, the problem of wispiness could be ameliorated by a sampling system or *in situ* analyser that could be towed through the ocean at a controlled depth.

The problem of choice of a tracer remained. Methane-21 would have to be measured mass spectrometrically. The feasibility of doing this measurement at sea was not known. As larger quantities would be needed than for ^{3}He it was not clear whether an injector like Ewart's could be used. Furthermore, we now know that methane-21 would be oxidized too rapidly in the sea to be used as a tracer (H. Craig, Fall 1982 AGU meeting).

The elimination of heavy methane as a tracer left perflourocarbons as the remaining viable choice. Discussions with J. Lovelock, A. Watson, and M. Whitfield in England convinced Broecker and J. Lupton that halogenated compounds detected by electron capture were the answer. It was also decided to follow the suggestion of P. Liss that we try to diffuse the tracer into the sea through permeable tubing rather than via the more cumbersome method used for Rhodamine dye. As the English group was also keen to do ocean experiments the American group decided work cooperatively with them on tracer technology.

PRESENT STATE OF DEVELOPMENT

Sulfur hexaflouride has been chosen as a tracer for the first experiments. Chromatographs are being set up and calibrated at the Marine Biological Associates' Laboratory in England and at Lamont to measure low levels of SF_6 in sea water. Watson is planning to measure the backgroud concentration in deep water samples this summer while the Lamont group is planning to test the ability to perfuse SF_6 through various types of tubing at pressures encountered in the ocean. Lupton at U.C.S.B. is studying the merits of tracer experiments in ocean basins and looking at the engineering aspects of seagoing injection and sampling systems.

The English group, under Lovelock, is continuing to evaluate other tracers. In principle there is a whole suite of halogenated sulfur compounds and halocarbons that could be used. Of particular interest are nonvolatile compounds which will not be lost to the atmosphere from surface waters, yet which are biochemically inert in seawater. Several volatile and nonvolatile tracers could be available within a few years.

Other projects which must be undertaken are studies of long term stability in seawater, development of efficient stripping methods for very low level samples, and development of sampling systems which are able to integrate over tens of kilometers to ameliorate the wispiness problem mentioned in the previous section.

WOCE should encourage these developmental projects and the preliminary mesoscale experiments planned for the near future.

PROTOTYPE EXPERIMENT

Advance toward large scale ocean experiments must be made in stages. One project under consideration would be focused on a 100 Km scale

experiment to be done in 1985-86, probably in conjunction with a Gulf Stream warm core ring study. The experiment will involve injection of a tracer within a few meters of a target isopycnal surface at a depth of a few hundred meters, followed by surveillance of the dispersion and mean vertical motion of the tracer for several weeks. Such an experiment will have many of the features of the large scale experiments sketched below. Injection, sampling and analysis systems will be tested under realistic conditions, but in compressed temporal and spatial scales.

During 1986-87 the engineering for a large scale injection could be completed so that by the time WOCE commences, the necessary technology for large scale tracer experiments could be ready.

OCEAN CIRCULATION EXPERIMENTS

Halogenated compounds and the electron capture detector enable studies of ocean circulation and mixing processes with length scales of order 1,000 to 10,000 Km and time scales of order 10 years. Water samples with concentrations of just 10^6 molecules/cm^3 can probably be analyzed with 95% confidence levels of ±5% every two or three minutes with a single *in situ* or shipboard gas chromatograph. Thus 10^{28} molecules (a few tons) could be accurately mapped after dispersal over an ocean volume of 10^{22} cm^3 (e.g. 3,000 Km X 3,000 Km X 1,000 m). The number of ocean circulation experiments possible with such capability is virtually unlimited.

For example, the western boundary undercurrent in the North Atlantic could be tagged in the north temperate zone. Mapping the tracer distribution over the succeeding 10 years would yield measures of the morphology of the undercurrent, the current speeds, the cross-isopycnal mixing in the vicinity of the current, and horizontal mixing both near and

away from the current. Although inadvertent transient tracers such as the freons may yield similar information to some extent, a purposeful tracer has the great advantage that its initial distribution will be accurately known. Furthermore the initial conditions may be chosen so that information from the purposeful tracer compliments information from the inadvertent tracers. Such a study would be fundamental to understanding the deep ocean circulation.

Another fundamental study could be carried out in a more quiet eastern basin. Tracer could be injected within ±10 meters of a target isopycnal surface at approximately 3000 meters depth. Assume an initial disc of tracer with a radius of 100 Km and a thickness of 20 m. Assume an horizontal eddy diffusivity of 10^7 cm^2/s, a vertical eddy diffusivity of 1 cm^2/s, and a mean vertical velocity of 4 m/year. After 10 years in such an admittedly oversimplified system, the distribution will have grown to a characteristic radius of order 1000 Km and a thickness of order 500 m. The 40 meter ascent of the centroid of the distribution might be measurable if care is used. Such an experiment would answer the long standing question of whether there is significant upwelling in the deep ocean interior and the related question of how large the cross-isopycnal mixing is in the interior.

As a final example, consider a study of cross-isopycnal mixing in the main thermocline of the North Atlantic. The early purposeful tracer meetings concluded that such an experiment was the most desirable first large scale experiment. The goal would be not only to measure the effective cross-isopycnal diffusion coefficient but also to distinguish mixing processes occurring where the isopycnal surfaces outcrop in the mixed

layer from processes occurring at depth. To this end both a volatile and a nonvolatile tracer would be injected in a layer a few meters thick centered about a chosen isopycnal surface at appoximately 500 meters depth in the Sargasso Sea. Volatile tracer which mixes along isopycnal surfaces up to the mixed layer at high latitudes will be lost to the atmosphere with a time constant of about one month. Nonvolatile tracer will not be so lost and may resubmerge at its leisure on different isopycnal surfaces, the potential density of its associated water having been modified by mixing, heating, and salinization in the mixed layer. Processes at depth, on the other hand, will affect the tracers as though they were indistinguishable. Thus the relative role of the outcrop regions in cross-isopycnal mixing of thermocline waters can be assessed. Of course, general horizontal mixing, and cross-isopycnal mean motion and mixing, in the thermocline region can also be studied in this experiment just as for the deep basin experiment sketched above.

The three experiments sketched here give just a taste of the possibilities offered by the use of halogenated compounds as purposeful tracers. Such experiments will prove an essential ingredient in future studies of the world ocean circulation. WOCE should support the effort for tracer development during the next few years in whatever way possible. WOCE should give serious consideration to including purposeful tracer experiments in its own program and should be prepared to help experiments launched outside the direct auspices of the program.

A CONCEPT OF WOCE

James C. McWilliams
August 8, 1983

1. INTRODUCTION

This essay is approximately the text of a talk given during the opening session of a WOCE workshop held at Woods Hole on the above date; it was one among four such talks presenting concepts of what a WOCE might be. It presents some personal opinions about relative priorities among possible techniques of observation and about dynamical concepts for the general circulation. I am not attempting to be comprehensive, neither on these particular topics nor on all aspects of a WOCE, since that is somehow to be the responsibility of the workshop as a whole.

My definition of the General Circulation is the distribution of and physics governing currents and related water properties on the horizontal and time scales within the following range:

$$\text{MESOSCALE} \lesssim (L, T) \lesssim (\text{GLOBAL, 5- OR 10-YEAR MEAN}).$$

In my opinion any finer restriction of the range of scales would be physically artificial and would lead to substantially less understandable results.

Some observing techniques can span this full range of scales; satellite systems are particularly suited to this. Inevitably, most in situ systems can only look at portions of this range. Obviously, a WOCE should place more emphasis on those which can see a greater portion. Also, however, there will be important, locally concentrated measurements at "pulse points" of the General Circulation---boundary currents, water property transformation zones, etc. In general, I believe a WOCE should include a complicated mix of measurements, and its design will be correspondingly complicated.

2. THE CENTRAL ELEMENTS OF WOCE AND THEIR RELATIVE PRIORITIES

My primary premise in considering the design of a WOCE is that fundamentally it should focus on circulation (i.e., velocity), on a variety of scales, and velocity measurements are generally a better basis for inferring other quantities of dynamical interest than are other observables, although the strength of this advantage obviously varies with the quantity and scale of interest. A secondary premise is that a WOCE can be done, which would be an enormous advance on our knowledge of the General Circulation, making use of both instruments and numerical models which either now exist or are under development with a reasonable expectation of fruition within a few years. The real challenge, in my opinion, is not to determine the feasibility of a useful WOCE, but rather to select among the possibilities and find the time, energy, people, and money to apply them.

A list of the central elements of a WOCE is presented in Table 1, ordered in three priority classes. Any such list is obviously premature and overly simplistic at the present

preliminary stage of planning for a WOCE, but it should be at least provocative, if not instructive, to describe the rationale behind this particular list.

Table 1
CENTRAL ELEMENTS OF WOCE

	Physical quantity	Principal measurement technique(s)
PRIORITY 1	surface geopotential	altimetry + tide gauges
	wind forcing	scatterometry; sea level pressure + meteorological model
	mid-depth velocity	deep drifters
PRIORITY 2	surface layer thermodynamics and circulation:	
	*buoyancy forcing	satellites (at least for short-wave radiation); ships; surface drifters; meteorological model
	*heat content	surface drifters; XBTs; [tomography?]
	*surface velocity	surface drifters; acoustic logs
	*mixing penetration depth	ships; drifters
	water-mass volumes and geography (i.e., "mean" lapse rate, dynamic height, chemical tracers; potential vorticity)	shipboard hydrography and water sampling
PRIORITY 3	variability of dynamic height, tracers, and interior heat content	shipboard surveys; tomography; purposeful tracer releases
	long-path velocity and area-averaged relative vorticity	tomography with reciprocal shooting; deep drifters (heavily averaged)

+ a variety of more local measurements (as required)

The first priority class comprises my concept of a minimal WOCE. Without all of these elements, I would be reluctant to defend the experiment as encompassing a sufficient portion of the presently unknown aspects of the General Circulation to be worth the bother of a WOCE. Of course, any of these elements would be valuable for a variety of oceanographic purposes, but less so than a WOCE. Altimetry and scatterometry have the right sampling characteristics to span the full range of General Circulation scales, as defined above. The wind forcing might also be obtained

by a complementary technique, a network of sea-level pressure measurements which would be converted to wind forcing by a suitable meteorological model. Mid-depth velocity (i.e., below the thermocline) is the best complement to surface geopotential for a full-depth determination of the velocity field (see Sections 3 and 4), and deep drifters are the most likely technique for sampling at least approaching the desired range of scales.

The second priority class includes buoyancy forcing. It is often considered secondary in magnitude to wind forcing in driving the General Circulation, although this is certainly not true by the measure of potential energy in the oceans even if it is by kinetic energy. Buoyancy forcing is also more difficult to deal with than wind forcing---practically, because available techniques are not likely to be as relatively accurate for the net forcing, and conceptually, because its "transmissivity" to the General Circulation below the surface boundary layer is much more subtle than for momentum. For this latter reason, WOCE should address the surface layer as a whole rather than simply the surface buoyancy fluxes. The global water masses have historically been the most useful indicator of the largest scales of the General Circulation, and any significant gaps in the historical collection should be filled in WOCE. These quantities provide the environment for the circulation; tracer distributions provide contraints upon possible circulation patterns (we are not yet very good at spelling out these constraints, partly because sources, sinks, and boundary conditions for tracers are not very well known); mean dynamic heights shore up the altimetric measurements where they are least accurate (at zero frequency, where the geoid must be subtracted); and large-scale potential vorticity is closely related to useful dynamical concepts for the large-scale circulation (Section 5).

Variability in the water masses has been assigned to the third priority class for two reasons: it is very labor intensive to approach the desired sampling rate (although tomography can in principle do much better at this than shipboard measurements), and dynamic height is relatively inefficient, compared to the combination of surface geopotential and mid-depth velocity, in determining circulation (see Sections 3 and 4 below), except in regions where the vertical structure of currents is particularly complicated. Obviously an important practical distinction for WOCE---which I won't attempt here---is the scale boundary between more useful "mean" water properties and less useful and more difficult to measure "variability". An interesting possibility in this latter category, which might not be subject to the two disadvantages stated above, is one or many purposeful tracer releases. The path-averaged velocity measurements which might be achieved with reciprocal shooting tomography might yield the largest horizontal scale components of the deep circulation more accurately than averages of deep drifter trajectories, and a connected circuit of them might yield the area-averaged relative vorticity within the circuit. One can state several reservations about this information---we have as yet no experience interpreting such information; it is uncertain how much of the energy of the deep circulation is on very long horizontal scales;

it is unclear how usefully dynamical concepts of the large-scale can be related to relative vorticity, which is the usually neglected component of the large-scale potential vorticity (see Section 5); and it is unclear how fine a horizontal resolution (i.e., smallest path length) is feasible. However, if satisfactory answers to these present uncertainties can be found, this technique may be a very valuable one for WOCE.

The final entry in Table 1, more local measurements, is important for WOCE, because certain features of the General Circulation are quite localized, but this will not be addressed further in this talk.

Finally, I would remark that the priority classes in Table 1 are not intended as sequential limits to the physical quantities and measurement techniques to be included in a WOCE. I believe all of these quantities and techniques should be included. The priority classes rather are intended to indicate the relative amounts of effort that might go into the various alternatives.

3. THE RELATIVE VALUE OF VELOCITY AND GEOPOTENTIAL MEASUREMENTS IN DETERMINING THE CIRCULATION

Part of the rationale for the ranking in Table 1 is the relative value of velocity and geopotential measurements in making estimates of the circulation. We can rely on the geostrophic relationship,

$$\underset{\sim}{v} = \hat{e}_z \times \underset{\sim}{\nabla} \phi$$

which allows us to consider their information content as equivalent in the absence of measurement or sampling errors. However, in the

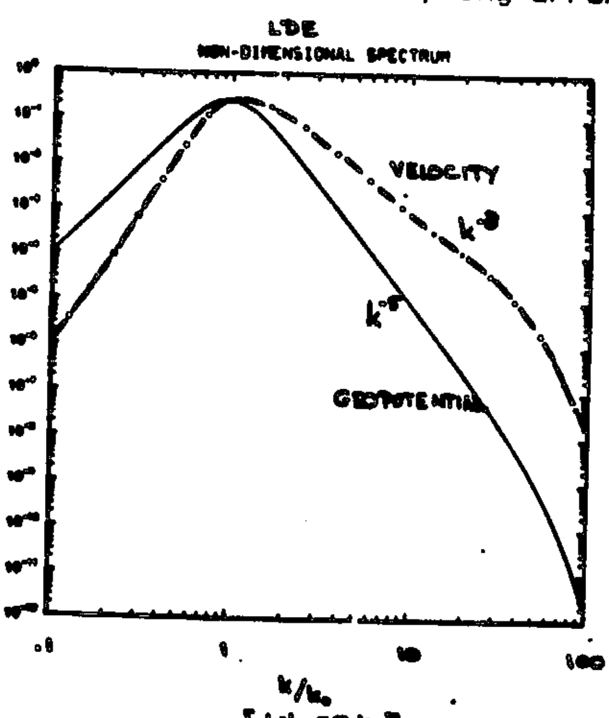


Fig. 1 (from McWilliams, Owens, and Hua, 1983)

presence of these practical deficiencies, the fact that, by geostrophy, the two fields must have different wavenumber spectra implies that geopotential measurements are relatively inefficient, compared to velocity measurements, in estimating velocity.

The differences in spectra are illustrated in Fig. 1, which is a statistical model fit to the observations in the POLYMODE Local Dynamics Experiment (LDE). Its empirical accuracy is least at the largest and smallest wavenumbers, where the spectrum shape has been chosen with theoretical prejudice, and greatest at intermediate wavenumbers. No wavenumber spectrum for the General Circulation is well known, of course, but this one from the LDE is among the best presently available.

A demonstration of the relative efficiency of the two types of measurements is the following. Consider a one-dimensional array of N data points, x_i, uniformly distributed over an interval of length L. At each of these points, either geopotential or velocity measurements are available. In either case we make optimal linear estimates of the velocity at any point, x, within the interval with the formulae

$$\hat{v}(x) = \sum_{i,j} \langle v(x)\,\phi(x_i)\rangle \left(\langle \phi(x_i)\,\phi(x_j)\rangle\right)^{-1} \phi_j$$

or

$$\hat{v}(x) = \sum_{i,j} \langle v(x)\,v(x_i)\rangle \left(\langle v(x_i)\,v(x_j)\rangle\right)^{-1} v_j ,$$

where the angle brackets denote ensenble mean covariances. The expected errors in such estimates are, respectively,

$$\langle (v-\hat{v})^2\rangle = \sum_{ij} \langle v(x)\,\phi(x_i)\rangle \left(\langle \phi(x_i)\,\phi(x_j)\rangle\right)^{-1} \langle \phi(x_j)\,v(x)\rangle$$

or

$$\langle (v-\hat{v})^2\rangle = \sum_{i,j} \langle v(x)\,v(x_i)\rangle \left(\langle v(x_i)\,v(x_j)\rangle\right)^{-1} \langle v(x_j)\,v(x)\rangle .$$

The required covariances are evaluated from

$$\langle \phi(x)\,\phi(y)\rangle = C(x-y)\left[1+\epsilon^2 \delta_{x,y}\right]$$

$$\langle v(x)\,\phi(y)\rangle = C'(x-y)$$

$$\langle v(x)\,v(y)\rangle = C''(x-y)\left[1+\epsilon^2 \delta_{x,y}\right],$$

where C(r) is the Bessel transform of the LDE geopotential spectrum (Fig.1),

$$C(r) = \int_0^\infty k\,J_0(kr)\,S(k)\,dk$$

and ϵ^2 is the relative error variance in a point measurement, modeled as white noise. For simplicity the same value for ϵ^2 is taken for each of the types of measurements. A summary measure of the estimation error is defined by

$$\langle \delta v^2\rangle / \langle v^2\rangle \equiv \int_0^L \frac{dx}{L} \langle (v-\hat{v})^2\rangle / \langle v^2\rangle ,$$

and this quantity is plotted in Fig. 2. The result is the anticipated one that, for a given N, errors from velocity measurements are smaller than those from geopotential measurements.

This result is reasonably robust in spectrum shape (except for very red shapes), the relative error level, and the interval

length. If the relative errors are sufficiently larger for the velocity measurements, of course, their competitive advantage can be reversed, although this is unlikely to be true in practice. If

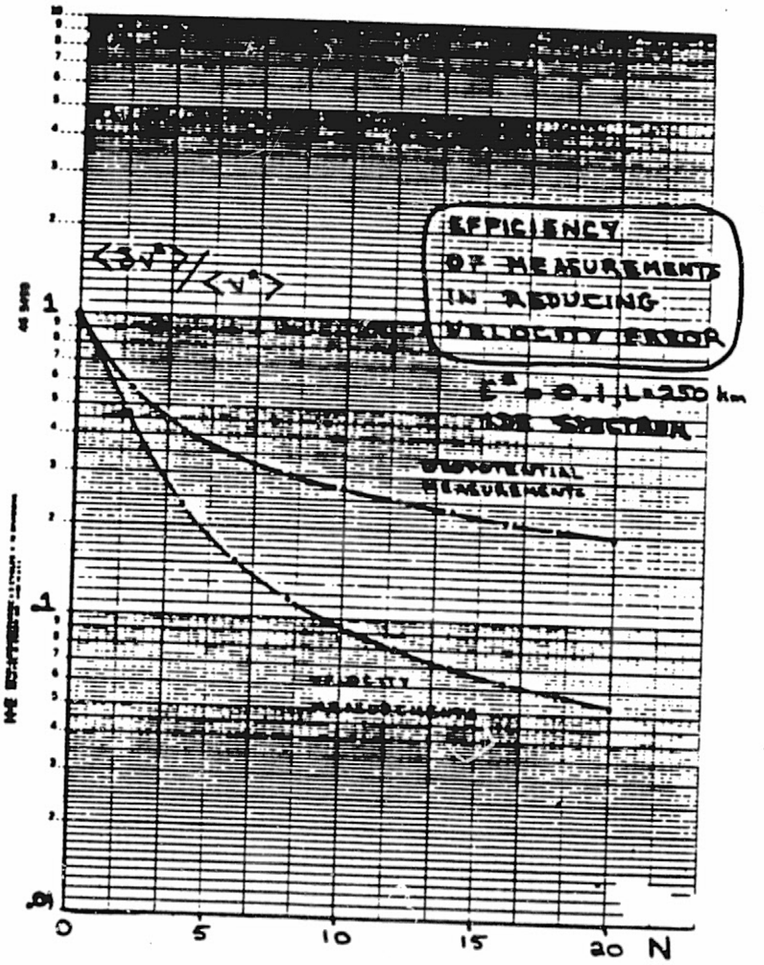


Fig. 2

alternatively we were to define the desired estimand as only the very large scale component of the velocity, then a competitive advantage would be given to geopotential measurements, opposing the tendency demonstrated above, since intermediate and small scale components of the spectrum in effect become errors in the estimation and the magnitude of this is larger for velocity than for geopotential.

In general, however, I believe the competitive advantage for WOCE will be found to most often lie with velocity measurements in estimating the circulation, and this is one of the reasons for giving dynamic height and temperature tomography measurements (both of which are closely related to geopotential) lower priority than mid-depth velocity measurements in Table 1. The same criticism, of course, can be applied to altimetric measurements of the surface geopotential, except that there is no

serious alternative for obtaining surface geostrophic velocity at the desired sampling rates.

4. VERTICAL SAMPLING ISSUES

Another reason for the previous assignment of priorities is due to the distribution of currents in the vertical. Consider a truncated modal representation for geostrophic velocity of the form

$$\underset{\sim}{v}(\underset{\sim}{x},z,t) = \sum_{n=1}^{M} \underset{\sim}{V}_n(\underset{\sim}{x},t)\, f_n(z) + \delta \underset{\sim}{V}(\underset{\sim}{x},z,t),$$

where the $f_n(z)$ are the familiar modes from linear theory, the $\underset{\sim}{V}_n$ are the modal amplitudes, and $\delta \underset{\sim}{V}$ is the residual associated with truncation after M modes. We model $\delta \underset{\sim}{V}$ as uncorrelated with any of the modal amplitudes and with itself except at zero lag. It has been shown, from long time series from moored current meters in several locations, that this representation is often an efficient one even with M = 2. Two examples of this are shown in Table 2; note that the depth-averaged relative error variances are quite small. The two sites have quite different modal statistics---the LDE has about three times the amplitude of MODE and its degree of cross-modal coupling is significantly less---associated with their different mean circulation environments (the LDE is within the Gulf Stream Recirculation Zone, while MODE has much weaker mean currents). The modes themselves vary slightly between the two locations, due to differences in the mean lapse rate, but this variation is secondary.

Table 2
MODAL STATISTICS (M = 2)

Quantity	MODE	LDE
$\langle \underset{\sim}{V}_n \cdot \underset{\sim}{V}_m \rangle$ $[\mathrm{cm}^2\,\mathrm{s}^{-2}]$	$\begin{array}{c\|cc} & m=1 & 2 \\ \hline n=1 & 19 & \\ 2 & 7 & 13 \end{array}$	$\begin{array}{\|cc} \hline 57 & \\ 10 & 21 \end{array}$
$\varepsilon^2 = \dfrac{\int_{-H}^{0} \frac{dz}{H} \langle \delta \underset{\sim}{V}^2 \rangle}{\sum_n \langle \underset{\sim}{V}_n^2 \rangle}$	.047	.054

With these modal statistics one can examine several aspects of different sampling schemes in the vertical. A very simple question is the degree of correlation between currents at different depths. With the modal representation, we write

$$\langle \underset{\sim}{v}(z_i)\cdot \underset{\sim}{v}(z_j) \rangle = \sum_{n,m} \langle \underset{\sim}{V}_n \cdot \underset{\sim}{V}_m \rangle f_n(z_i) f_m(z_j) \left[1 + \varepsilon^2 \delta_{z_i,z_j} \right],$$

and evaluate it for three levels: at the surface, in the thermocline, and at mid-depths. The results are shown in Table 3.

Table 3
CROSS-LEVEL VELOCITY CORRELATIONS

	MODE			LDE		
	0	-700	-2500	0	-700	-2500
z = 0 m	1.00			1.00		
-700	.94	1.00		.93	1.00	
-2500	.23	.33	1.00	.23	.40	1.00

They demonstrate the general result that surface and thermocline geostrophic currents are highly correlated, hence nearly redundant measurements, while deeper currents are only weakly correlated with the shallower ones, hence a nearly independent measurement.

Another question one can pose is this: suppose one has measurements of $\underset{\sim}{v}(z=0)$ and $\underset{\sim}{V}_2$---the first baroclinic modal amplitude---how well can one estimate $\underset{\sim}{v}(z)$? This question is an idealization of the possible combination of altimetric measurements of the surface flow with either dynamic height or tomographic temperature measurements of the baroclinic flow, disregarding here the inefficiency of converting geopotential to velocity information (Section 3). It also can be answered in the context of optimal linear estimation theory using the preceding modal representation and Table 2 statistics. The results are shown in Table 4. Errors in estimating $\underset{\sim}{v}(z)$ in and above the thermocline are satisfactorily small, whereas they are quite large below the thermocline. This and the preceding calculation clearly indicate the importance of a set of deep velocity measurements.

Table 4
$\underset{\sim}{v}(z)$ ESTIMATION ERRORS FROM $\underset{\sim}{v}(0)$ and $\underset{\sim}{V}_2$

z [m]	$\langle (\hat{\underset{\sim}{v}}(z)-\underset{\sim}{v}(z))^2\rangle / \langle \underset{\sim}{v}(z)^2\rangle$	
	MODE	LDE
-700	.10	.13
-2500	.53	.46

It is unclear how generally an M = 2 expansion is an efficient representation of the General Circulation, and geographical exploration of this issue would be very useful for the design of WOCE. It seems likely that this representation will be less successful for circulation on the largest scales than it has been shown to be for the point variability sampled by one-to-several-year moorings, although we cannot discount the possibility that some other low-order modal representation, perhaps with different modes, might be comparably efficient. In any event, if, as also seems likely, the largest scales of circulation are more surface intensified than the mesoscale and

intermediate scale currents which dominate the MODE and LDE statistics, then the necessity for deep velocity measurements will be even greater for adequate estimates of deep circulation.

Thus the arguments of both the preceding and present Sections tend to lower the priority accorded to the baroclinic geopotential in the context of altimetry and deep velocity. Some of the counter-arguments are listed in Section 2.

5. LARGE-SCALE DYNAMICS

At the risk of both oversimplification and as yet far too little empirical justification, I propose that the point dynamics of the large-scale, low-frequency (i.e., beyond mesoscale) General Circulation are, to a substantial degree, the advective-diffusive dynamics of stretching vorticity.

Potential vorticity can be split into three components for small Rossby number:

$$q = P + S + R,$$

where P represents the combined effects of Coriolis frequency variations and topography, S the vertical stretching of fluid columns, and R the relative vorticity. The first two of these are often expressed as f/h, where f is the Coriolis frequency and h is the thickness of a fluid layer bounded by surfaces of constant potential density or the ocean bottom. In this notation the mean potential vorticity equation is

$$\frac{\bar{D}}{\bar{D}t}\,(\,\bar{P} + \bar{S}\,) = -\nabla\cdot\overline{\mathbf{v}'S'} \quad + \quad \begin{array}{l}\text{wind forcing}\\ \text{buoyancy forcing}\\ \text{relative vorticity}\\ \text{sub-mesoscale effects}\\ \ldots\end{array}$$

Here the overbar denotes large-scale, the first term on the right is associated with mesoscale motions, and the list refers to other processes generally of lesser magnitude. This equation expresses what I mean by the dynamics of stretching vorticity. It is much more likely to be valid as a point balance, which is the form in which it is written, than in a large area or volume integral, where the "lesser" terms will rise in relative importance.

The accompanying heat equation is

$$\underset{\text{(advection)}}{\frac{\bar{D}}{\bar{D}t}\,\bar{T} \approx \bar{w}\frac{\partial\bar{T}}{\partial z}} = \underset{\text{(diffusion)}}{-\nabla\cdot\overline{\mathbf{v}'T'} + \ldots}$$

This balance has been demonstrated at the LDE site in the Gulf Stream Recirculation Zone, where both the mesoscale eddies and large-scale circulation are strong, and may carry-over to other, weaker regions. The inferred vertical velocity is plotted in Fig. 3. An indication of the dominance of stretching vorticity dynamics is the relative smallness of the Subtropical Gyre Ekman pumping (w_E) at the upper surface. Estimates of the strength of

mesoscale diffusion at several locations are shown in Fig. 4. It clearly has significant geographical variation, and in some locations is quite large. There appears to be a monotonic increase with mesoscale eddy energy, although there is no reason to believe that the straight-line fit has general validity.

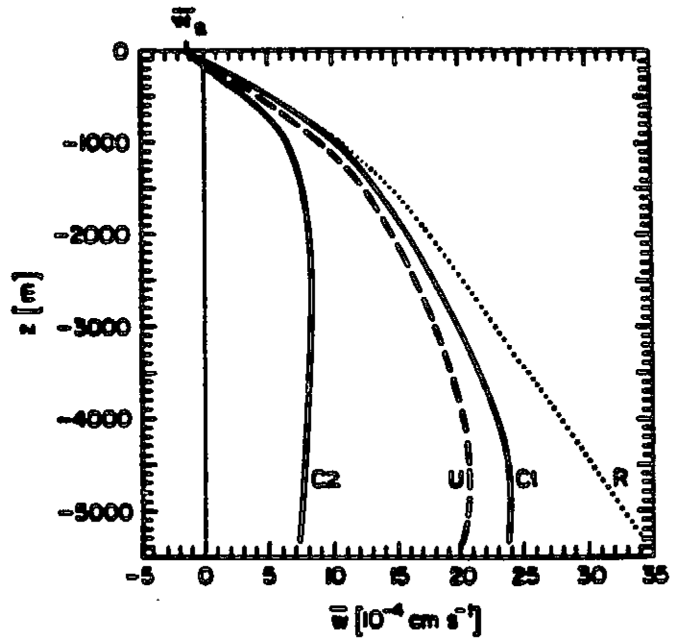


Fig. 3. Several inferential estimates of mean vertical velocity at the LDE site (McWilliams, 1983).

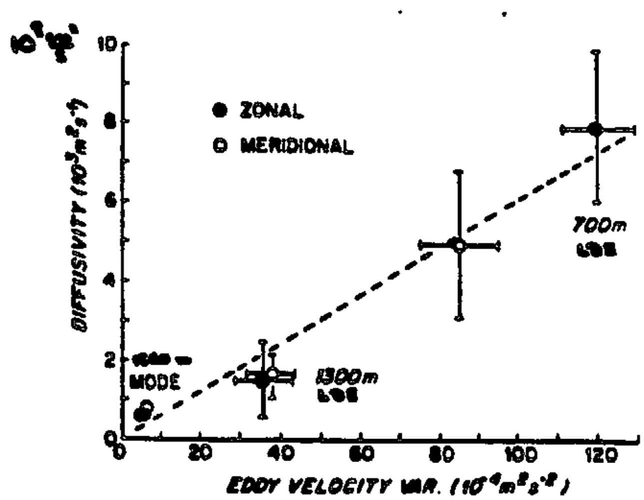


Fig. 4. Horizontal, single-particle diffusivity vs. the variance of eddy velocity (Price, 1983).

Several studies have recently focused upon the large-scale distribution of f/h. A particular interest in these studies has been the possibility of spatially extensive, homogeneous regions where this quantity has little variation. For several years now, eddy-resolving numerical-model solutions have exhibited homogeneous regions when the interior, sub-mesoscale diffusivities are sufficiently small. An illustration of f/h in the North Atlantic is shown in Fig. 5. In a thin layer in the upper thermocline, there is a region which might be homogeneous;

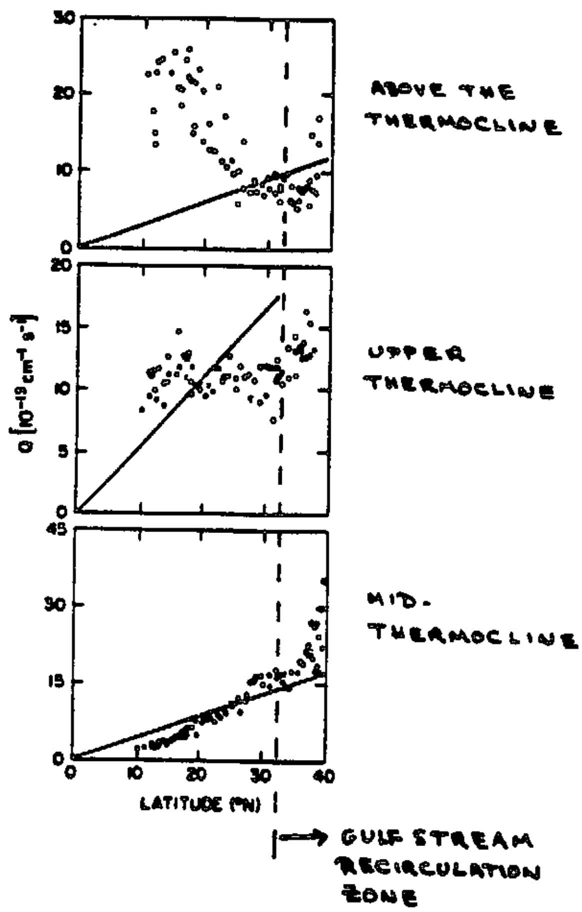


Fig. 5

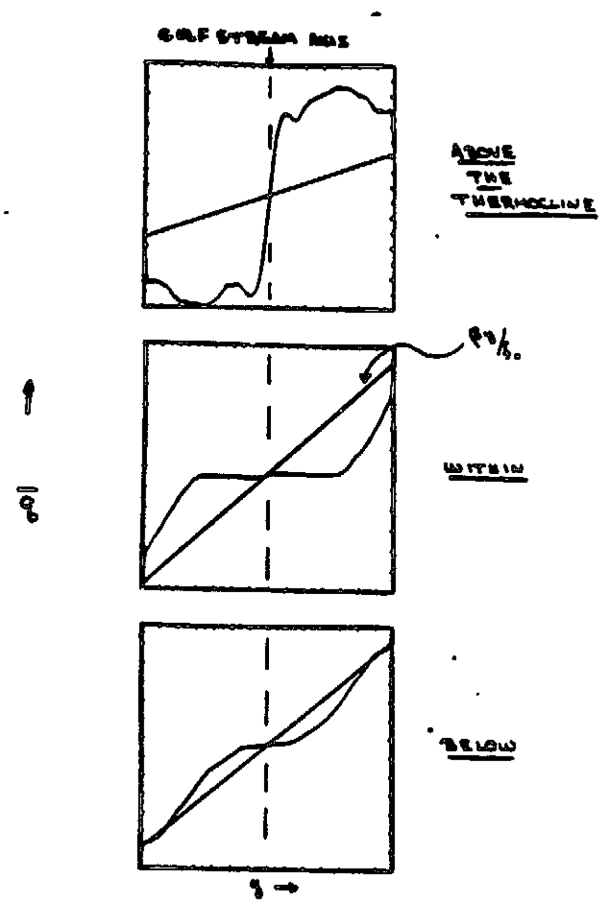


Fig. 6

Fig. 5. $\bar{P} + \bar{S}$ values along 65° and 50° W. Longitude (solid and open circles, respectively) for three layers bounded by different σ_θ surfaces (McDowell, Keffer, and Rhines, 1983). The line in each panel indicates the contribution from f variations alone.

Fig. 6. $\bar{P} + \bar{S} + \bar{R}$ profiles from a numerical model solution (Holland, personal communication).

certainly any systemmatic variations are weaker than the contribution from f alone. There is substantial scatter about a

possible uniform value which, if not either measurement noise or compensated by local relative vorticity, would suggest that the large-scale f/h is not exactly uniform (n. b., in a truly homogeneous region both the mesoscale and the large-scale q are uniform). In any event, the possibly homogeneous region ends at the outer boundary of the Gulf Stream Recirculation Zone.

An example of $\bar{\bar{q}}$ profiles from a two-gyre, numerical model solution is shown in Fig. 6 (in most locations one can neglect the difference between q and f/h). In this case there are extensive homogeneous regions, extending through the Gulf-Stream and its Recirculation Zones. This solution appears to be much more homogeneous than the North Atlantic. However, the North Pacific appears to be much more homogeneous than the North Atlantic (Keffer and Rhines, personal communication) and may therefore be better matched by this model solution.

I certainly hope that a WOCE will yield good empirical descriptions of the elements of stretching vorticity dynamics and the distribution of potential vorticity on the larger scales of the General Circulation, so that theoretical ideas on these topics can be better assessed.

6. DISCLAIMERS

There are many aspects of a WOCE which I have not talked about and are fully important as those I have---the reasons for wanting to do a WOCE, special regions, numerical modeling, tracer and water mass transformation dynamics, human and institutional arrangements needed for accomplishing a WOCE---and this neglect is only for lack of time.

Ocean Acoustic Tomography — Basin Scale Applications

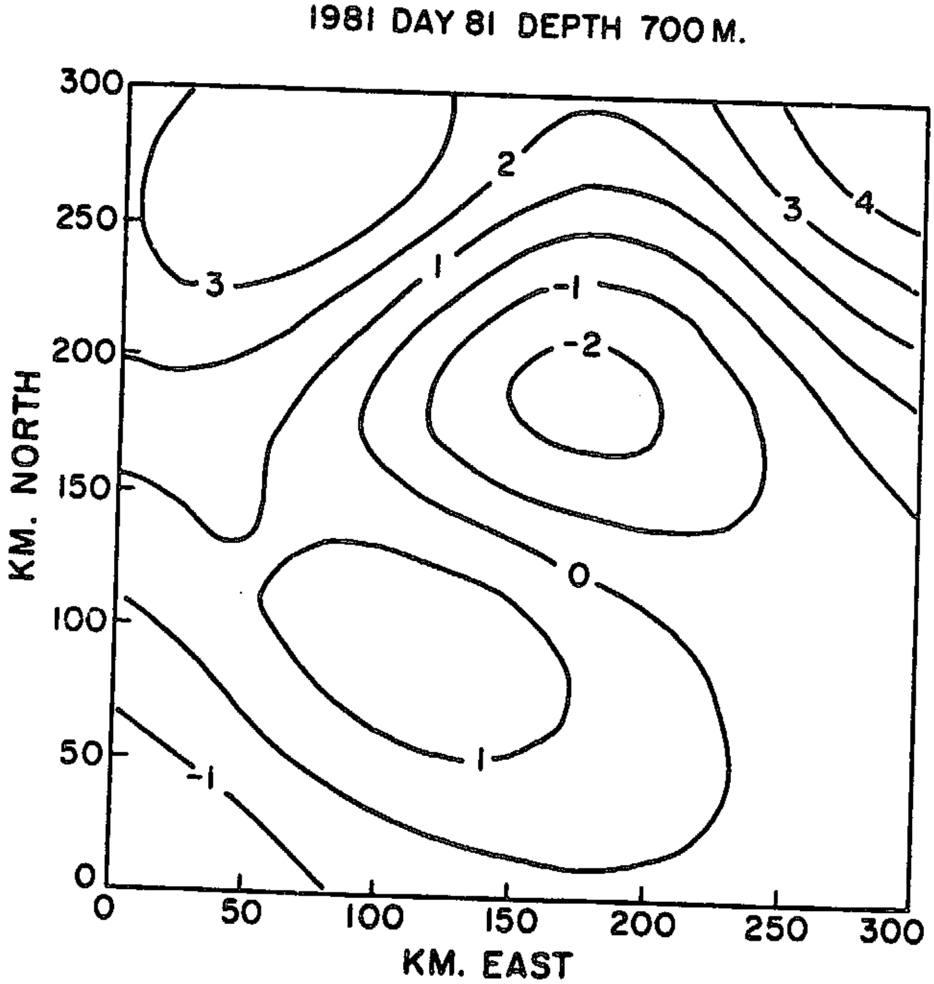
Walter Munk and Carl Wunsch

1. Introduction

Ocean acoustic tomography is an experimental technique for observing the ocean by passing known acoustical signals through the ocean and inverting the received signals for the oceanic state along the acoustic paths. Since it was first proposed (Munk and Wunsch, 1978), an intensive effort has ensued to demonstrate and understand its capabilities (e.g., Spiesberger *et al*., 1980; Spindel and Spiesberger, 1981).

The fullest demonstration of the practical capabilities of a tomographic system was provided by a field test in late 1981 and described in preliminary form by The Ocean Tomography Group, 1982, and in more detail by Cornuelle, 1983, Cornuelle and Wunsch, 1983, Spiesberger *et al*., 1983. Here a 300 km x 300 km square was mapped acoustically using 4 sources and 5 receivers. Figures 1-3 show three "snapshots" of the ocean (from Cornuelle, 1983); the point we wish to emphasize is that these maps were made *purely* acoustically - the only non-acoustic information being a single, average, hydrographic station.

This demonstration showed the ability of tomography to map the mesoscale. In contemplating the general ocean circulation, the most appealing property of tomography is not our ability to make a mesoscale map — rather it is the natural *integrating* behavior of the system. To zero order, the travel time of an acoustic pulse across an ocean basin is a measure of the integrated heat content of the ocean. The *difference* in travel time for a pulse to go from A to B and that from B to A is at zero order a measure of the average horizontal *velocity* in the vertical



Figs. 1-3. Sound speed in ocean at 700m. depth made from purely acoustic observations with instrumentation around the periphery. Sound speed is highly correlated with temperature. (From Cornuelle,1983). Such maps were constructed at 3 day intervals but only three are shown here. Other depths are also mappable.

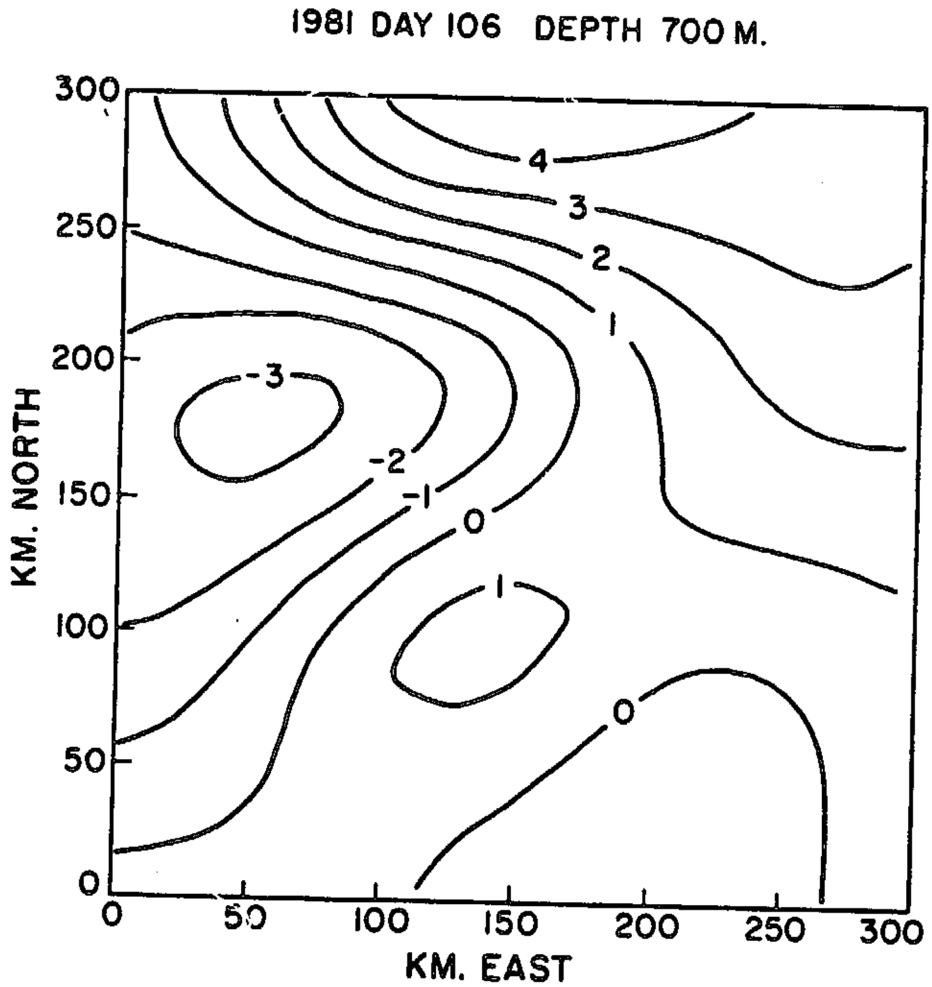

1981 DAY 106 DEPTH 700 M.
KM. NORTH
KM. EAST
0
50
100
150
200
250
300
4
3
2
1
0
-1
-2
-3

Fig. 2

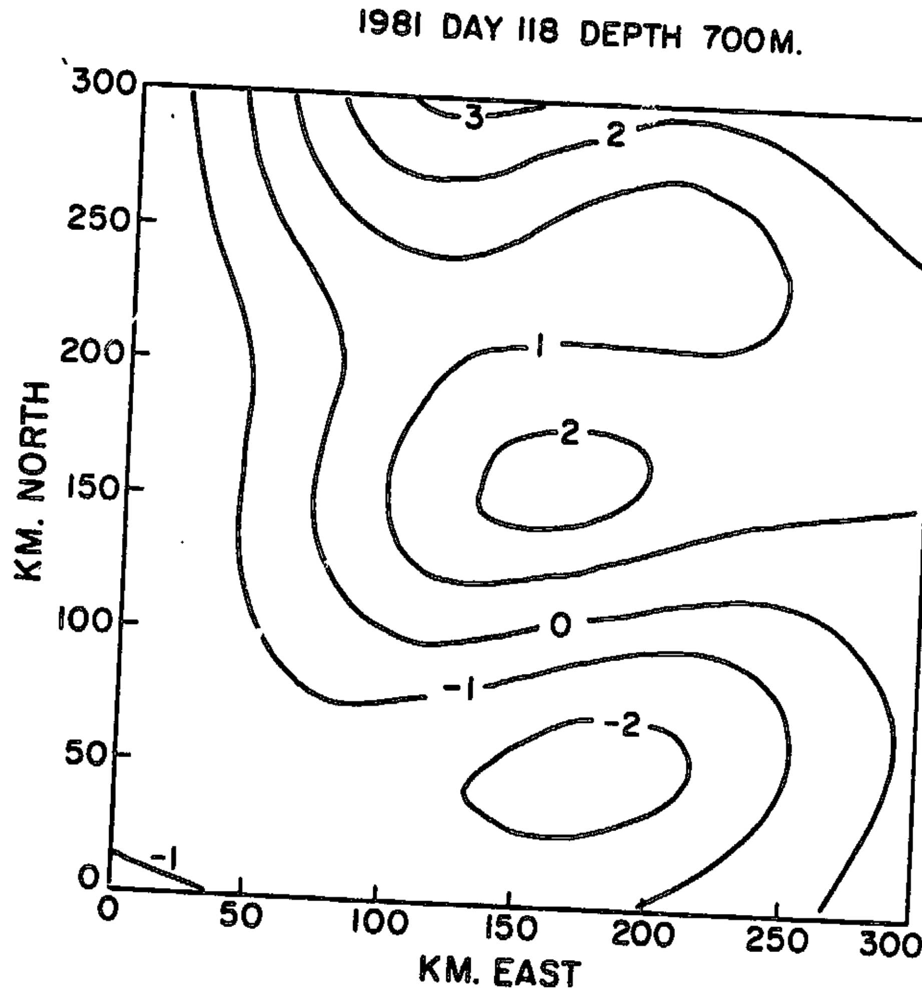

1981 DAY 118 DEPTH 700M.
KM. NORTH
KM. EAST
300
250
200
150
100
50
0
0
50
100
150
200
250
300
3
2
1
2
0
-1
-2
-1

Fig. 3.

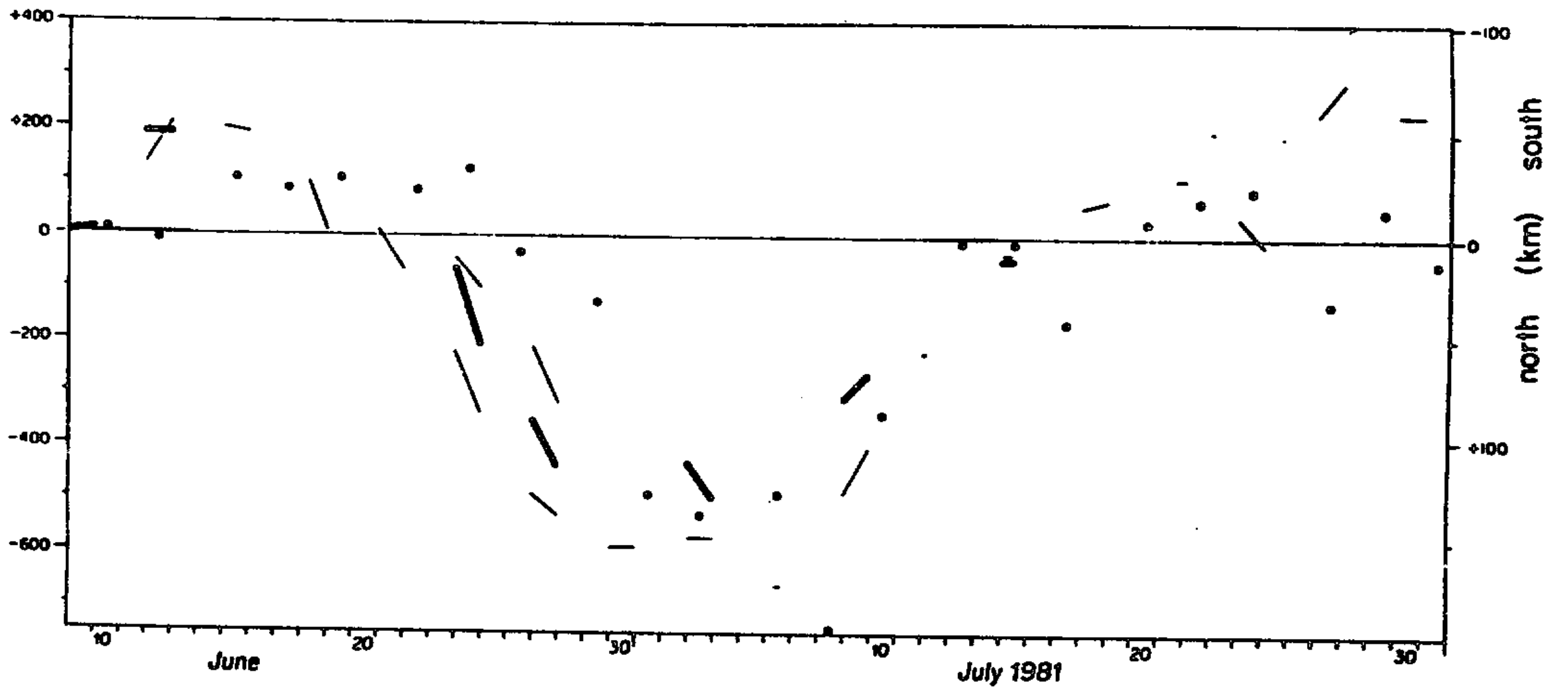


Fig. 4. Comparison of Gulf Stream position from IR and other conventional data (dots), with position inferred from tomographic transmission across it (lines). From Spiesberger, et al 1983.

plane of A-B. (One way travel time tomography — "temperature tomography" is what was demonstrated in 1981; reciprocal travel time tomography — velocity tomography — is being demonstrated this summer, 1983, at sea.)

Munk and Wunsch, 1982a, b considered the implications of these ideas at some length, including an error analysis of the system if one seeks the integrated heat content and oceanic vorticity. Excerpts from MW 1982a are enclosed as an Appendix. The conclusion was that in principle very precise determination of oceanic heat content, velocity, vorticity and upwelling could be made.

(Figure 4, taken from Spiesberger <u>et al</u>., 1983 shows one way travel times measured <u>across the Gulf Stream</u> perhaps one of the most difficult such places to make the acoustics work. The very large changes in acoustic travel time are a reflection of meanders of the Gulf Stream thus changing the relative amounts of warm Sargasso Sea water and cold slope water along the acoustic path. Details of the meander process are contained in the details of the changes in arrival time — but the zero order "climatological" effect is apparent even to the eye.)

2. <u>What Should We do with Tomography?</u>

We know of no other practical approach to obtaining large spatial averages of the ocean that would not require inordinate amounts of hardware at sea. Another tomographic virtue is that the growth in number of paths goes nearly quadratically with the number of instruments in the water — rather than the linear information growth characteristic of point measurements (Longuet-Higgins, 1983 has analyzed the information content of N source/receiver pairs).

Tomography is an example of an observational system, now under development, potentially capable of making crucial observations for WOCE and its climate objectives. The issue for WOCE planning is whether the promise of the technology is so great as to warrant an accelerated effort to make it nearly certain that the technology will indeed be available when needed, or whether existing efforts are perceived as adequate to do so, or whether it is such a speculative technique that it should remain only as an experimental component of WOCE.

Those of us intimately involved with acoustic tomography have certain prejudices. Based upon our experience to date, both with field data, and with theory, we see no barriers to making the method a potent component of WOCE by the end of the decade. Indeed the technology used in the 1981 experiment — that for temperature tomography — is essentially off-the-shelf hardware. Velocity tomography is somewhat more difficult — it requires a better clock; we believe that this technology will also be fully demonstrated within 2 or 3 years. By the beginning of the intensive field phase of WOCE (circa 1989) the following should have happened (under present planning):

1) Demonstration of reciprocal tomography over 300 km (1983) and over order 2000 km (1986).

2) Demonstration of temperature tomography over 300 and 1000 km (1981, 1983, 1986).

3) Simulation (Rizzoli) of basin scale measurements of both types of tomography in an EGCM.

4) Complete understanding of mid-ocean accuracies and precisions of the measurement.

5) An ability to construct and deploy comparatively modest numbers of moorings in ocean basins (about 15/basin) see Fig. 5 capable of generating strong integral constraints on the circulation and its fluctuations.

6) Data relay by satellite to remove the necessity of (a) mooring recovery for data recovery, (b) "blind" waiting to know if instruments are operating.

7) Development (principally by the U. S. Navy) of techniques adequate for tomography from moving ships (relying heavily upon the very high navigation accuracies to be made available from the GPS).

If we can in fact do these things, then as outlined by Munk and Wunsch, 1982a, the combination of altimetry, scatterometry, tomography and models should vastly increase knowledge of what the ocean is doing.

<u>The crucial elements as seen by us are the following</u>:
The forthcoming demonstrations must be clear and convincing; the ability to communicate via satellite is essential; commercial manufacture of instruments for both types of tomography must begin very soon. These things can all be done — 1989 is probably the <u>earliest</u> one could hope to deploy significant numbers of tomographic moorings. It seems worth a good effort.

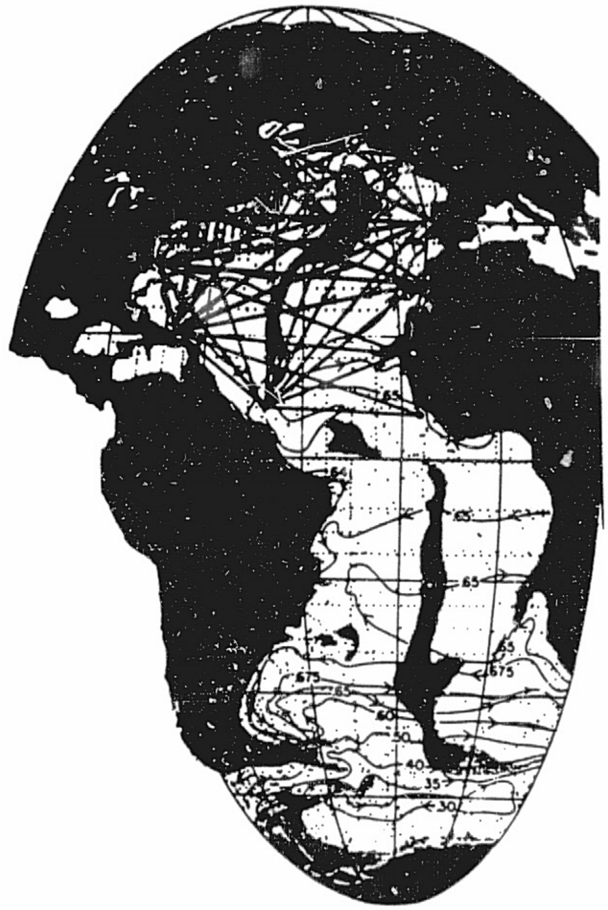


Fig. 5. Conceivable scope of a 15 mooring North Atlantic gyre scale measuring array. Only a few of the 105 total paths are drawn. Along each, one would measure the integrated heat content and velocity as a function of depth, as well as circulation around each closed set of paths.

APPENDIX

(From Munk and Wunsch, 1982a)

4.1. *Temperature*

An analysis of the errors in a temperature measurement is fairly straightforward and is given in appendix A. The conclusion is that changes by 10^{-3} K (1 mK) in the average along a 2500 km path could be detected. At 2500 km distances, we expect to be able to resolve a minimum of 15 ray paths, corresponding to a vertical resolution of 100 m at the base of the mixed layer (Munk & Wunsch 1982*a*) and 1000 m in the abyssal ocean.

Thus we can map the mean heat content in the box as a function of depth and time. Because of the large dimensions of the box, the 'noise' associated with mesoscale eddies is greatly subdued. Upper and lower bounds to the heat content can be computed with the extremal methods of inversion described by Wunsch & Minster (1982). These procedures provide a stable estimate of the bounds, given the vector of measurement error.

4.3. *Currents*

Our goal is to measure travel time to one part in 10^6, corresponding to measuring currents to 10^{-6} the speed of sound, or $1.5\ \mathrm{mm\,s^{-1}}$. Standard current meters have precisions of $1\ \mathrm{cm\,s^{-1}}$. However, the principal asset is not the precision but the large-scale spatial integration, and this leads us to the next topic.

4.4. *Vorticity*

Perhaps the most challenging application of the velocity measurement is to determine large-scale vorticity. This was suggested by Rossby (1975) but to our knowledge has never been done. By measuring the 'sing-around' travel times along the periphery of the proposed array, clockwise and anticlockwise, and forming the difference, one gets a direct measure of the total vorticity within the array. This can be done for each of the four interior triangles (three of which are independent) to obtain some information about vorticity gradients. One elegant feature is that the need for precise time-keeping is dispensed with in a sing-around geometry (appendix C).

Because tomography has the ability to resolve the ocean in the vertical, the vertical stucture of the mean vorticity can be combined with the vertical structure of the density field to yield potential vorticity as a function of depth. This type of information may provide very powerful constraints on analytical and numerical investigations of oceanic circulation. For reasons already stated in § 3.4, most modern theory is in terms of vorticity as the independent variable.

How accurately can we measure vorticity? Consider an $L \times L$ square. From Stokes theorem the circulation $4Lv$ equals $L^2\zeta$ where ζ is the mean vorticity in the square, and v is the mean velocity around the periphery. Thus $\zeta = 4v/L$. But $v/C = \frac{1}{2}\delta\Delta/\Delta$ where $\delta\Delta$ is now the difference in anti-clockwise minus clockwise sing-around time. We have given a somewhat optimistic argument that $\delta\Delta/\Delta$ can be measured to a precision of 10^{-6} independent of L (see appendix A). This corresponds to a vorticity

$$\zeta = (2C/L)(\delta\Delta/\Delta) = (1.2 \times 10^{-3})\,10^{-6} = 1 \times 10^{-9}\,\mathrm{s}^{-1} = 1 \times 10^{-5} f$$

for $L = 2500$ km and $f = 2\Omega \sin(\text{latitude}) \approx 10^{-4}\,\mathrm{s}^{-1}$. The required mooring position-keeping is $10^{-6}L = 2.5$ m, and $\delta\Delta = 1.7$ ms.

What is the expected signal? In a Sverdrup balance

$$\rho\beta v h = \operatorname{curl}\tau,$$

where ρ is density, h is layer depth, $\beta = \mathrm{d}f/\mathrm{d}y$ (y is northward) and τ is the wind stress. Set $v = \partial_x\psi$, $\zeta = \nabla^2\psi$, $\beta = f/r_{\text{Earth}}$. Let τ vary from $-\tau_0$ in the trades to $+\tau_0$ in the westerlies over a distance L. The resulting gyre vorticity is

$$\zeta_g = \nabla^2\psi = \psi_{yy} = 8x r_{\text{Earth}}\tau_0/\rho L^3 hf = 10^{-8}\,\mathrm{s}^{-1} = 10^{-4}f$$

for $x = 3000$ km designating the distance from the eastern boundary, $L = 2500$ km, $h = 1$ km, $r_{\text{Earth}} = 6000$ km, $\tau_0 = 1\,\mathrm{dyn\,cm^{-2}}$ $(10^{-5}\,\mathrm{N\,cm^{-2}})$, and $f = 10^{-4}\,\mathrm{s}^{-1}$. This is ten times our precision estimate.

The mean vorticity in a mesoscale eddy might be $\zeta_e = 10^{-1}f$, and so be enormous compared with the gyre vorticity $10^{-4}f$. However, with $L_e = 100$ km and $L_g = 2500$ km, the eddy integrated vorticity $L_e^2\zeta_e$ might be about the same as the gyre $L_g^2\zeta_g$, and we can expect a significant fluctuation in the measured circulation as mesoscale eddies enter and exit the array area.

For an array consisting of n peripheral transducers, Longuet-Higgins (1982) has shown that $n(n-1)/1 \times 2$ is the number of paths between instrument pairs, $n(n-1)(n-2)/1 \times 2 \times 3$ is the number of triangles, and $(n-1)(n-2)/1 \times 2$ is the number of independent triangles. For a square array (figure 2) there are six travel paths (and thus twelve independent reciprocal transmissions) and four triangles, but only three of the triangular sing-arounds are independent (since $\Delta_{1231} + \Delta_{1341} = \Delta_{12341} = \Delta_{4124} + \Delta_{4234}$). For a pentagonal array we get six independent measures of vorticity, enough to determine the coefficients in a power-series expansion (with subscripts designating differentiation)

$$\zeta = a + b\zeta_x + c\zeta_y + d\zeta_{xx} + e\zeta_{xy} + f\zeta_{yy},$$

so one might consider a vorticity balance

$$(\partial_t + v\cdot\nabla - A\nabla^2)\,\zeta = \operatorname{curl}\tau,$$

with the terms on the left-hand side acoustically measured, except for the diffusivity A.

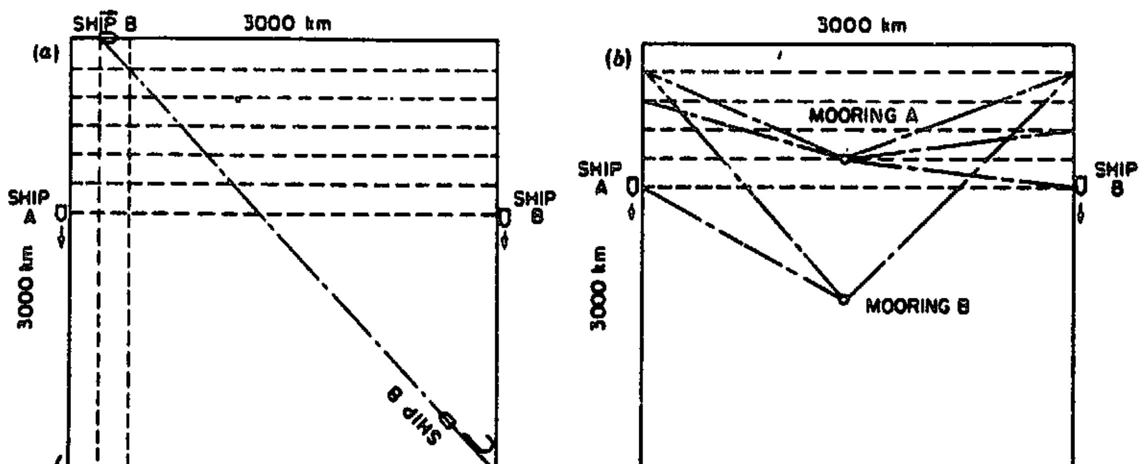


FIGURE 5. Ship-to-ship tomography over a large area. This requires very precise navigation, as may become available from the Global Positioning System in the mid-1980s. For two ships stopping every 50 km for 30 min (this may not be necessary), a tomographic coverage with 100 km resolution is completed in 13 days (a). The combined ship-mooring tomography (b) increases the number of transmission paths from one to six, and greatly reduces the coverage time.

4.5. *Upwelling*

The ability to measure velocities and densities over moderate-sized regions of the central ocean permits a connection to be made to ordinary vorticity dynamics and leads to an ability to infer vertical velocities. The condition of non-divergence states that

$$u_x + v_y + w_z = 0.$$

Conservation of density requires that

$$u\rho_x + v\rho_y + w\rho_z = 0.$$

Thus with an array like *B* of figure 3 we can estimate w or w_z. In practice one would solve for both w and w_z by inverting the two equations simultaneously. We estimate that w as small as $10^{-5}\,\mathrm{cm\,s^{-1}}$ (*ca.* $1\,\mathrm{m\,d^{-1}}$) would be measurable, a value representative of open sea upwelling.

The density conservation equation can be written

$$-w = u(\rho_x/\rho_z) + v(\rho_y/\rho_z) = -uh_x - vh_y,$$

where h_x and h_y are the slopes of isopycnal surfaces (in the notation of Stommel & Schott 1977). Combining this with the linear quasi-geostrophic momentum balance equations leads to the beta-spiral equation

$$uh_{xz} + v(h_y - \beta z/f)_z = 0.$$

All these terms can be measured, at least in principle, and so the beta spiral provides a set of consistency relations that are useful in the inversions.

4.6. *Further developments*

As the Global Positioning System (GPS) (see Klepczynski 1978) becomes operational, the possibility for doing acoustic tomography from ships is opened up. The GPS level of accuracy in navigation (*ca.* 1 m horizontally) would remove the ship position error from the measurement, provided that towed sensors (perhaps arrays) can be navigated relative to the ship with high accuracy. One could envisage ship-to-ship tomography (figure 5*a*) or a mixed ship-to-mooring tomography (figure 5*b*), permitting rapid mapping of large ocean areas. This would eliminate clock and power problems, but with the present arrangement a high processing gain is achieved by many repetitions of the coded sequence in a single transmission; travel-time precision and path stability is attained by averaging many such transmissions. There is a question of how all this can be done in a moving geometry.

Appendix A. Precision and resolution at megametre acoustic ranges

The measurement error of arrival time for a resolved ray path is $(S/N)^{-\frac{1}{2}}\Delta$, where S and N denote signal and noise power, and Δ is essentially the reciprocal of the radiated bandwidth. Write $\Delta = Qf^{-1}$, where Q is the conventional 'Q' for a resonant source, and f is the centre frequency. For a 300 km transmission experiment in the fall of 1982 we have $Q = 4$, $f = 400\,\mathrm{Hz}$, hence $\Delta = 0.01\,\mathrm{s}$, and $S/N = 20\,\mathrm{dB}$, thus giving a measurement error of 1 ms, and a fractional error of 6×10^{-6}. Can this precision be retained at megametre ranges?

We wish to minimize $(S/N)^{-\frac{1}{2}}\Delta$. The signal intensity S is attenuated by α dB/km, with $\alpha \approx 10^{-7}(f/\mathrm{Hz})^2 \equiv \alpha_0 f^2$ below a few hundred hertz. This can be written $S/S_0 = \mathrm{e}^{-\beta R}$, with $\beta = \frac{1}{10}\alpha(\ln 10)$. At high frequency $S^{-\frac{1}{2}}$ is larger, and at low frequency Δ is larger (assuming constant Q). A minimum in $S^{-\frac{1}{2}}\Delta$ corresponds to a maximum in

$$S\Delta^{-2} \approx f^2 \exp\{-\tfrac{1}{10}(\alpha_0 R \ln 10) f^2\},$$

which yields an optimum frequency $f_{\mathrm{opt}} = (10/\alpha_0 R \ln 10)^{\frac{1}{2}} = 208\,\mathrm{Hz}$ at $R = 1000\,\mathrm{km}$. Note that

TABLE A1. ACOUSTIC TRANSMISSION AT 224 Hz OVER 1 Mm

(1) source level	189 dB/μPa	an array of sources at the same total level would yield a directivity gain of 3–6 dB
(2) spherical spreading	−120 dB	cylindrical spreading beyond 10 km gives −100 dB; our experience indicates that spherical spreading is more nearly correct.
(3) attenuation	−6 dB	
(4) receiver directivity	+4 dB	present configurations of four receivers
(5) coherent averaging	+16 dB	mean of 38 sequences
(6) pulse compression	+24 dB	for 225 digits per 8 s sequence, for a 32 Hz bandwidth†
	+107 dB	
(7) noise per hertz	−68 dB	medium shipping density at sea state 3
(8) bandwidth	−15 dB	for 32 Hz bandwidth†
(9) signal:noise	24 dB	
digit length	32 ms	for 32 Hz bandwidth
precision	1 ms	

† Bandwidth cancels between lines (6) and (8).

$\beta R = 1$ for $f = f_{opt}$, so that the absorption is always by e^{-1} at the optimum frequency. Cylindrical spreading would reduce the intensity according to $S \approx R^{-1}$, and so the dependence of $S^{-\frac{1}{2}}\Delta$ on range (but always at optimum frequency) is

$$S^{-\frac{1}{2}}\Delta \approx R^{\frac{1}{2}} f_{opt}^{-1} \approx R^{\frac{1}{2}} R^{\frac{1}{2}}.$$

The *fractional* error in travel time is independent of range. (All this is specific to a chemical absorption $\alpha \approx f^2$.) We have ignored the frequency dependence of the noise $N(f)$; at a few hundred hertz this is dominated by ship noise.

The conclusion is that the fractional travel time error (which is insensitive to range) is roughly 6×10^{-4}. Table A1 summarizes the expected performance. The table is based on conservative estimates. The precision can be further improved by using sources of greater power and shorter digit length, and the range extended from 1 to 3 Mm. For comparison, the fractional change in soundspeed is 3×10^{-6}/mK (for temperature tomography), and $7 \times 10^{-6}/(\text{cm s}^{-1})$ (for velocity tomography).

Additional errors are introduced by timekeeping and mooring position keeping. Referring now to the experimental setup described by the Ocean Tomography Group (1982), clock error can be kept to 1 ms by use of a rubidium oscillator, and mooring motion can be monitored to 1 metre (corresponding again to 1 ms) by interrogating bottom transponders. These are acceptable errors.

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A WOCE Density Program

W. D. Nowlin, Jr. and Joseph L. Reid

1. Introduction

The focus of this discussion is on the relationship between the World Ocean Circulation Experiment (WOCE) and measurements of the ocean's density field and its variations. The questions addressed here include: What is the present status of relative dynamics fields for the world ocean; What can a density program contribute to a WOCE; Can WOCE measurements aid in interpretations of the density field; and What are reasonable elements of a WOCE density program?

The purpose here is not to provide answers to these questions, but rather to identify some problems involved with planning a density program for the study of the world ocean circulation and thereby to provide a framework for further study which may yield more definitive solutions to such problems.

2. The global mean ocean circulation

Our current picture of the mean global ocean circulation is almost exclusively derived from historical hydrography (here meaning serial measurements of temperature, salinity and pressure at ocean stations sufficient to determine the density field for geostrophic flow estimations) and chemical measurements (ranging from dissolved oxygen to radionuclides).

Many specifics of the general ocean circulation are still not yet well described, much less understood. For example, until the past few years the Antarctic Circumpolar Current was thought of as a broad, sluggish eastward drift around Antarctica. We have now learned that the Antarctic Circumpolar Current, the major zonal current system in terms of transport, is really comprised of several narrow current cores with much attendant mesoscale, energetic behavior. This knowledge was gained by new field observations -- of the density and property fields, as well as the velocity fields. As another example, it appears that we know so little about zonal circulations at subsurface depths -- we have only the vaguest qualitative picture -- that even the best analysts or modelers might be led astray by the present data bank of principally zonal sections. Some yet unanswered questions about the North Atlantic are: Does the Gulf Stream (or North Atlantic Current) really extend east of the Mid-Atlantic Ridge at depths down to 3000 m? And does the Gulf Stream recirculation begin in the eastern half of the Atlantic? How far east? And what is the real circulation that accounts for the pattern of the Mediterranean salt tongue? (Even a few meridional sections placed properly might do much to resolve these issues.)

Moreover, our picture of the general circulation, even in well sampled regions, may not be exactly accurate since knowledge of the density field provides us only with flow at one level relative to another.

If we can measure the slope of the sea surface, or of any other pressure surface, with respect to a level surface, we can calculate the geostrophic current that would balance this slope, or horizontal pressure gradient. But we are not yet able to locate a level surface in the ocean.

We can calculate, using the hydrostatic equation, the slope of an isobar with respect to another isobar, and then use this slope to calculate the geostrophic current at one of the isobars with respect to the other. But this is relative, not absolute, geostrophic speed.

If we assume that the flow in the deep ocean is very weak, i.e., much less than at the sea surface, then the flow at the sea surface calculated relative to some very deep isobar may not be far wrong, and we can learn something about the general circulation of the upper waters.

This has been the general practice among oceanographers and is sometimes called the dynamic method. Following it we have learned something about upper circulation and vertical shear (relative dynamics), though our results cannot be as exact as we would like.

We have known for a long time that there are deep currents in the ocean. That means that a deep isobar does not coincide with a level surface, but slopes. The surface slopes we have calculated will be incorrect by the magnitude of the slope, and the surface speeds will be too strong or too weak by the magnitude of the deep flow along our reference isobar. Likewise, the reverse is true; that is, if we take as known by other measurements the slope of the surface isobar, then the slope of deep isobars will be in error by the error of the surface slope measurement.

When representing the surface or intermediate flow fields, we commonly resort to dynamics maps relative to some deep reference level. In fact this is the only means available at present for describing the flow at one level relative to another in considering large areas. This leveling of the sea surface or some intermediate depth isobar with respect to a deep reference cannot be extended to the global ocean for several reasons. Bathymetric features -- the mid-ocean ridge system and constraints such as Drake Passage -- separate the basins of the world ocean at intermediate depths. The mediterranean seas, such as the Arctic, are separated from the world ocean by rather shallow sills. Also, the sea surface area over continental shelves is large, but cannot be leveled accurately relative to deep reference surfaces.

3. Relative dynamics fields for the world ocean

Lisitzin's (1965) map of the height of the sea surface with respect to the 4000-decibar (db) surface (Fig. 1), which lies at depths near 3850 m, is a familiar type of relative dynamics display. Though the detail is not as good as later versions, it illustrates the range of values quite well. The highest values are near the tropic circles, within the great anticyclonic gyres, and the lowest values that we can relate to the others are found on the southern side of the Antarctic Circumpolar Current. The greatest range is about 270 cm, from the 400-cm height just south of Japan to the 130-cm value within the Weddell Sea. However, the 400-cm contour is inside a ridge and should not be included, so instead we have a global range of 380 - 130 = 250 cm. (Note that the North Pacific values are higher than those in the North Atlantic by about a meter on this map. This may be a little too high, but not much.)

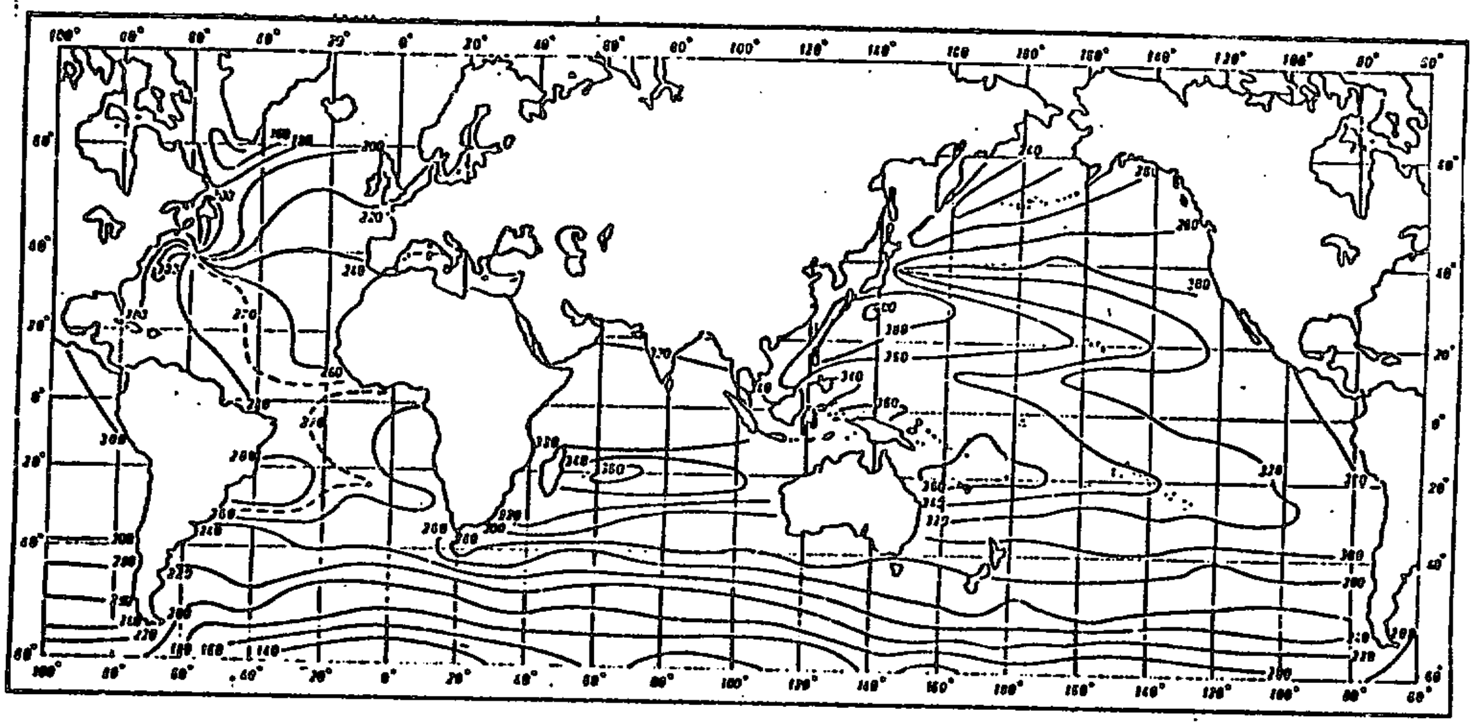

Fig. 1. The distribution of different heights of mean sea level (dyn cm) in the world ocean (Lisitzin, 1965).

The greatest slopes occur across the western boundary currents -- the Gulf Stream and Kuroshio -- and are about 100 cm in 100 km, or 10^{-5}. However, the greatest increments from a high to a low are across the Antarctic Circumpolar Current -- about 150 cm in 2000 km, or 10^{-6} in slope.

A later version for the Pacific is given as Fig. 2 (Reid and Arthur, 1975). Here the sea surface is referred to only 1000 decibars, which allows more detail, but does not give the full range. We see a range of 188 cm from 220 cm south of Japan to 32 cm in the Antarctic -- a little less than the 200 cm range relative to 4000 decibars. This means, of course, that the range of relative slopes below 1000 decibars is much smaller, and much greater accuracy in measurement is required to describe the geostrophic shear in these deeper waters.

The newer data, integrated from 4000 decibars, show the range to be 268 cm -- from 378 off Japan to 110 in the Ross Sea. However, there is a ridge rising above 4000 decibars in between, which makes the real range uncertain. North of the ridge the range from 15°S, 170°E (360 cm) to 62°S, 170°W (163 cm) is only 197 cm.

Fig. 3 shows the sea surface relative to 2000 decibars for the Atlantic (Reid et al., 1977). The slope is roughly the same as for the Pacific, with a range of about 170 cm from the high near Cape Hatteras to the low in the Weddell Sea. This is a little less than the slope of the sea surface relative to 1000 db in the Pacific, which was 188 cm.

Referred to 4000 decibars there is a range of about 200 cm in the Atlantic, compared to 273 in the Pacific. Most of the difference is between the northern highs: the Pacific high is 88 cm higher than the Atlantic high, but the low is only 16 cm higher.

Fig. 4 gives Wyrtki's (1971) map of the sea surface relative to 3000 decibars, showing many analogous features.

The difference in dynamic heights between the Atlantic and Pacific has been discussed by Reid (1961). Between 25°N and 25°S the sea surface in the Pacific stands 25-50 cm higher than does the Atlantic relative to 1000 db. The Atlantic waters are simply much denser than the Pacific because the salt content is greater.

The difference in height between the two oceans can be compared by geodetic leveling across the Isthmus of Panama. These maps show that there is a slope upward to the west along the equator in the Atlantic and Pacific (though not in the Indian Ocean), so we are comparing a relative low in the Pacific to a relative high in the Atlantic. Leveling with theodolites across the Isthmus finds the Pacific about 23 cm higher. Oceanic leveling by using the 2000 decibar surface as a reference gives about 16 cm. (We note there are seasonal and other variations that our oceanic data bank cannot yet take into account.) The agreement, though encouraging, should not be taken as evidence that oceanic leveling around the world, based on deep isobaric surfaces, is that good.

One reason for doubt is that the oceans are separated by ridges as well as continents. Not only does the mid-ocean ridge system separate the major basins, but the Bering Strait (at about 60-m depth) and the Greenland-Scotland Ridge (at about 800-m depth) make leveling into the Arctic Mediterranean very dubious.

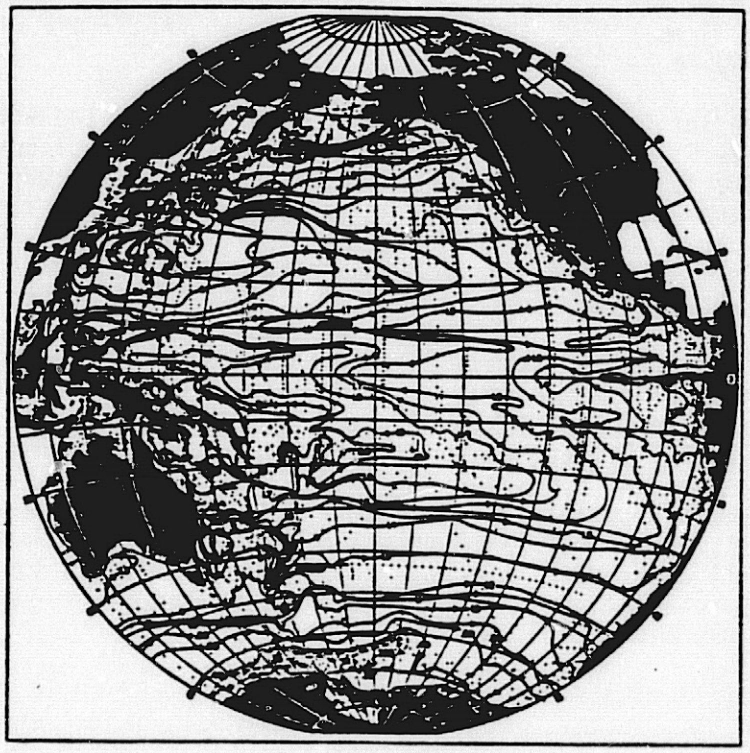

Fig. 2. The geopotential anomaly at the sea-surface relative to the 1000-decibar surface, in dynamic meters ($10\ m^2/sec^2$ or $10\ J/kg$). In the shaded areas the ocean depth is less than 1000 m. (Reid and Arthur, 1975).

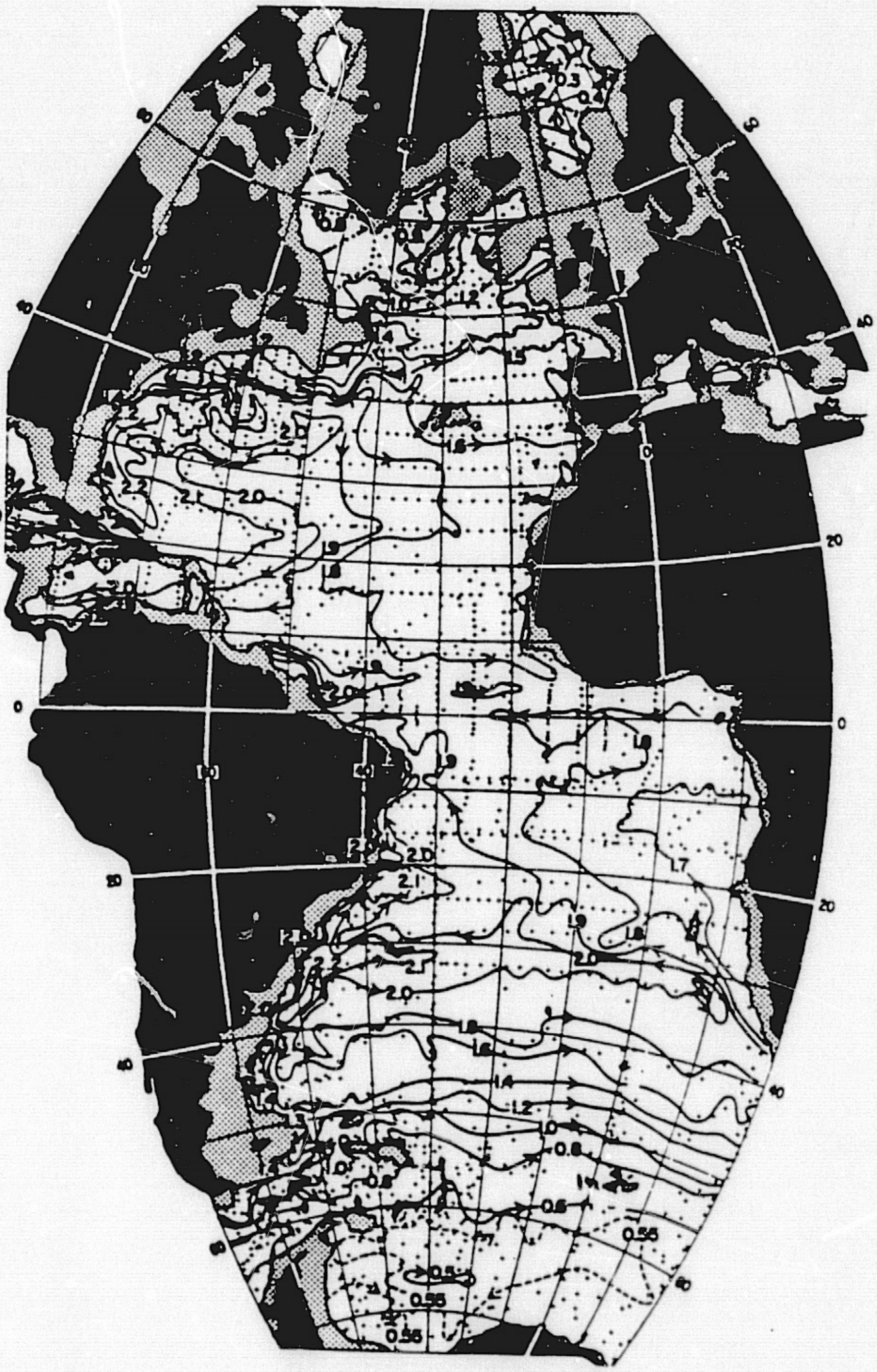

Fig. 3. The geopotential anomaly at the sea surface with respect to 2000 db in dynamic meters (10 J kg^{-1}). (Reid, Nowlin and Patzert, 1977, figure 8).

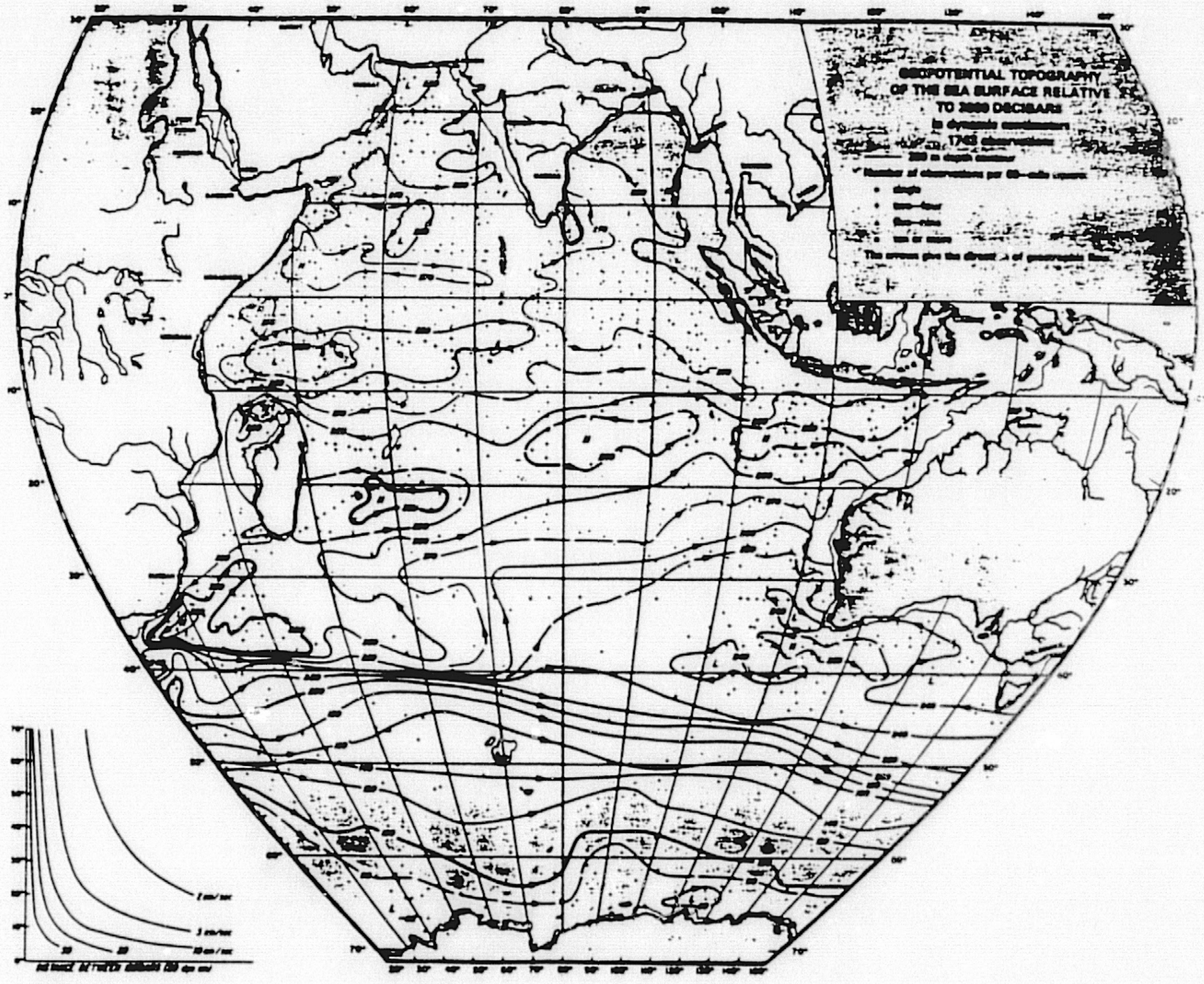

Fig. 4. After Wyrtki (1971).

Constrictions between the major basins, such as Drake Passage, may also invalidate any oceanic leveling referred to deep isobars.

We conjecture that the flow from the Bering Sea into the Arctic Ocean through the Bering Strait is a consequence of the higher sea level in the Pacific than the Atlantic, but we have not a valid measure of the height of the Arctic Ocean. A measuring system that allows estimation of the elevation of the Arctic and Norwegian-Greenland Sea with the world ocean would be very valuable. It would have to extend to at least 70°N.

While most of the great ocean is open at 2000-db depth, it is plain from Fig. 5 (Reid, 1981), which shows the slope of the 2000-db surface with respect to the 3500-db surface, that one of these surfaces certainly slopes. While the ranges in each ocean north of 40°S are small, perhaps 5 cm, there is still an ocean-to-ocean difference of about 5 cm in the north. South of 40°S the effect of the Antarctic Circumpolar Current, shows a difference of about 35 cm between 40°S and 60°S.

From Fig. 6 (Reid and Arthur, 1975) we see that in the far south Pacific there is still a range of about 10 cm in 1000 km at 3000 decibars relative to 4000 decibars. This figure also emphasizes that oceanic leveling referred to a deep isobar is limited to reference depths of less than 4000 m.

There are also problems when attempting to carry oceanic leveling into shallow water. For example, measurements of currents and of sea level at tide gauges suggest there are slopes of 10 to 20 cm over the last 50 km near the shore in the northeast Pacific, i.e., slopes of 10^{-6}. Using these techniques we are unable to incorporate flow representations for shallow regions into global pictures relative to deep reference levels.

There are many regions in the world ocean in which the density field is not well sampled, even for the construction of major circulation features. The workers who prepared the relative dynamics maps shown previously have of course faced that problem and have come up with their interpretations. A different scientist dealing with the same materials might have produced an alternative interpretation. This is not to say that the overall field of relative dynamics would necessarily change, but the details would likely be different. And, some of these details are very important features.

East-west sections were made in the South Atlantic aboard the METEOR and again during the IGY. However, the zonal flows in the eastern South Atlantic have never been mapped in detail. Wüst (1935), Defant (1941a, 1941b), Kirwan (1963), Reid (1981) and Fu (1981) each have worked through the data and have, with their various methods, tried to understand the flow. However, with the materials available, no one has been able to achieve satisfactory results. Fu (1981) applied inverse methods to these east-west sections, but perhaps because of the orientation of the lines, he discussed only the meridional flow of Wüst's (1935) Antarctic Intermediate Water and North Atlantic Deep Water across the lines, and did not deal with the zonal flow patterns of these waters, or with any waters below the North Atlantic Deep Water. As his transport streamlines (Fig. 7) indicate, the constraints of the 8°-latitude spacing of the lines precluded a satisfactory treament of flow parallel to those sections. The same limitations have plagued previous investigators and are equally bothersome on the map (Fig. 5) of steric height at 2000 relative to 2500 db prepared by Reid (1981.

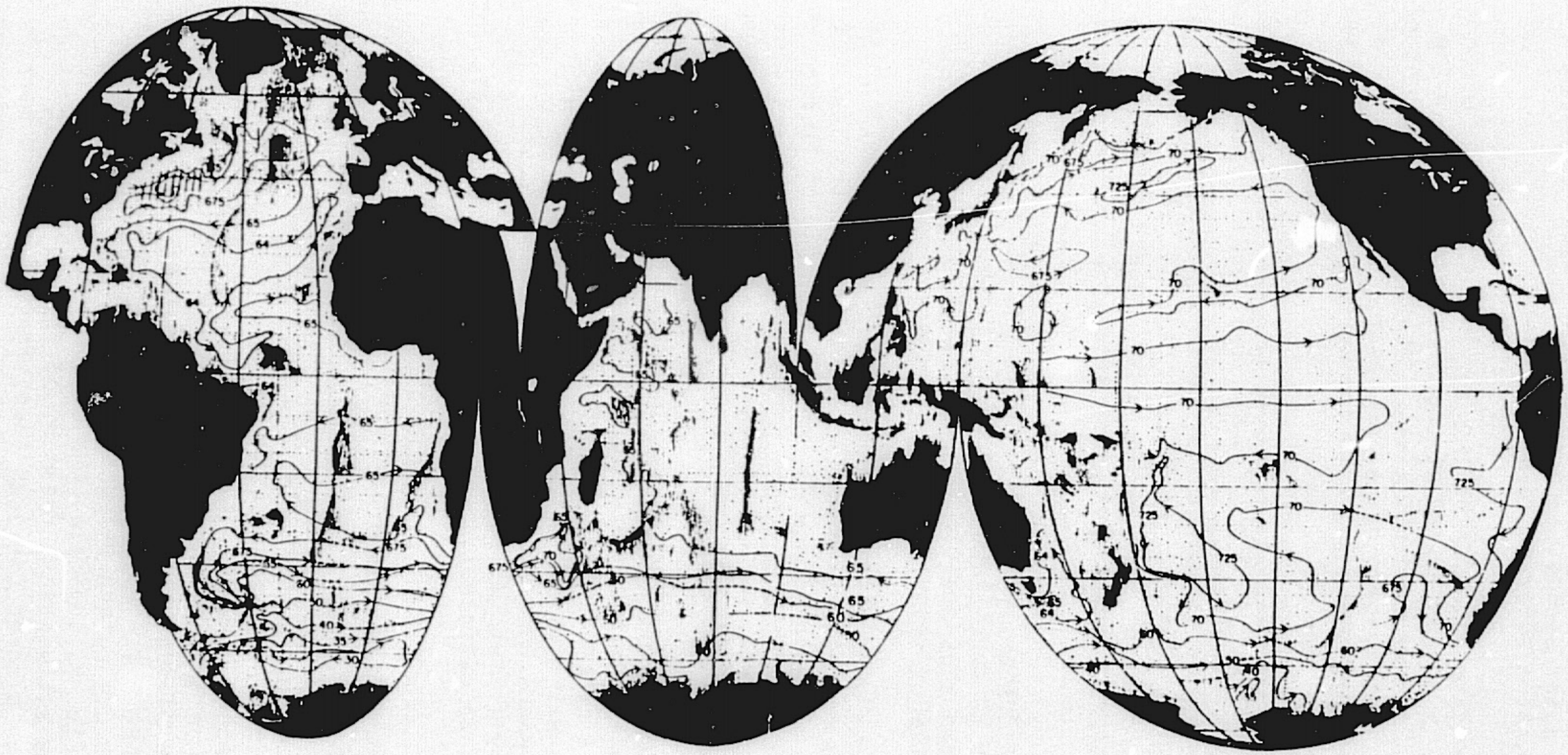

Fig. 5. Geopotential anomaly of 2000-db surface relative to 3500 db, in dynamic meters. Shaded area is less than 3500 m depth (Reid, 1981).

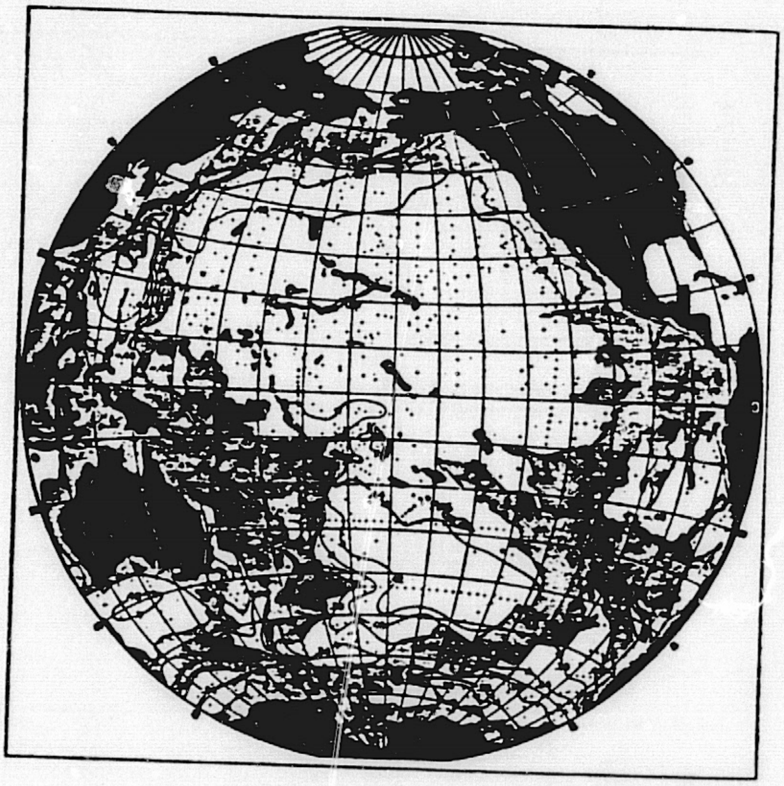

Fig. 6. The geopotential anomaly at the 3000-decibar surface relative to the 4000-decibar surface, in dynamic meters ($10\ m^2/sec^2$ or 10 J/kg). In the shaded areas the ocean depth is less than 4000 m. (Reid and Arthur, 1975).

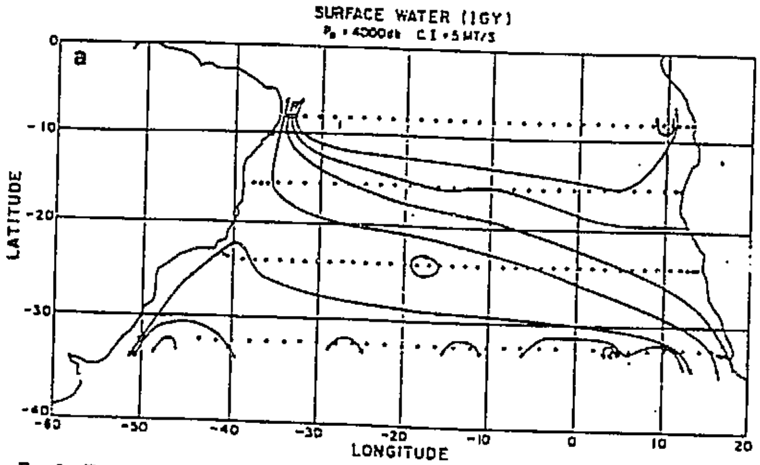

FIG. 9a. Transport streamlines for the Surface Water of the IGY sections with p_e = 4000 db. Contour interval (C.I.) is as indicated in the figure heading.

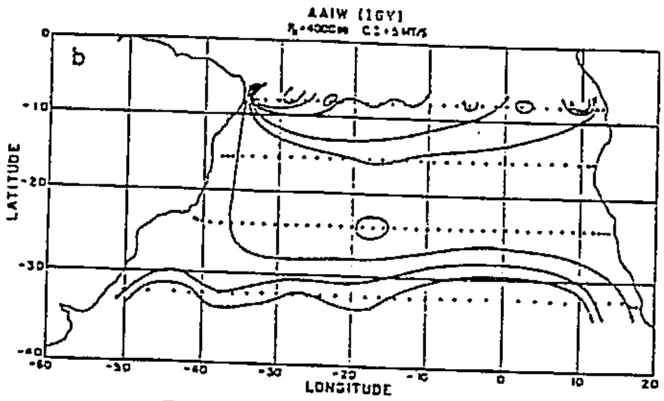

FIG. 9b. As in Fig. 9a except for the AAIW.

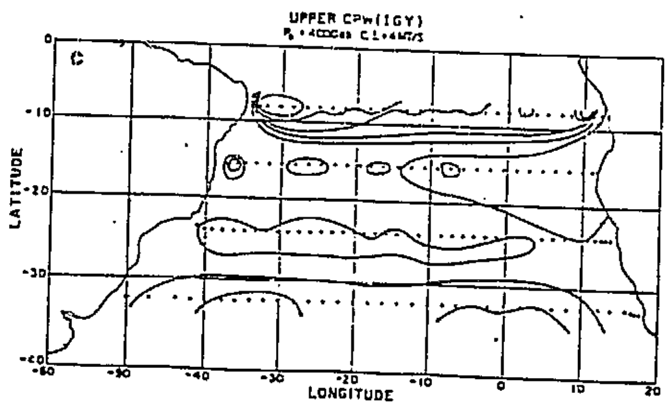

FIG. 9c. As in Fig. 9a except for the Upper CPW.

Fig. 7. Transport streamlines for six water layers in the mid-latitude South Atlantic as determined through application of inverse methods to IGY data by Fu (1981, fig. 9).

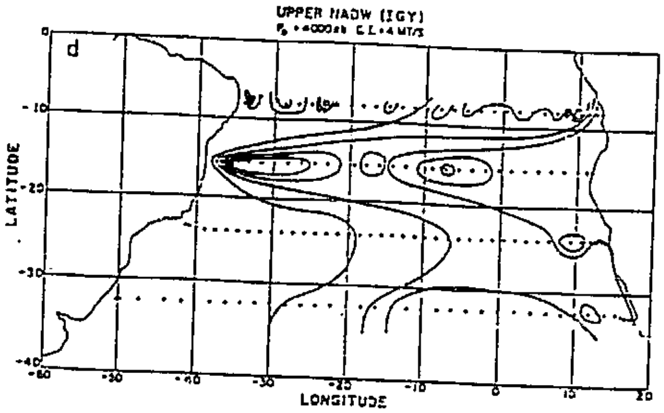

FIG. 9d. As in Fig. 9a except for the Upper NADW.

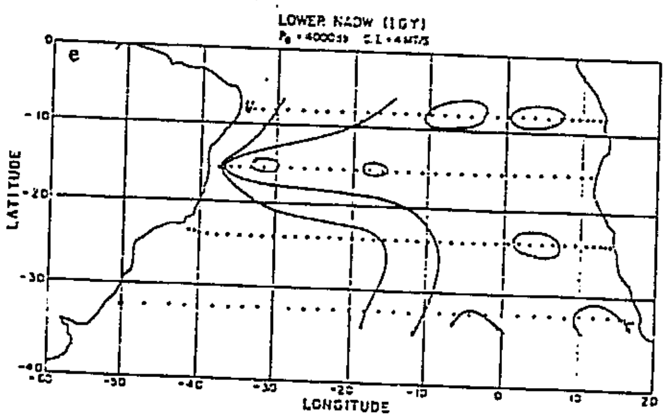

FIG. 9e. As in Fig. 9a except for the Lower NADW.

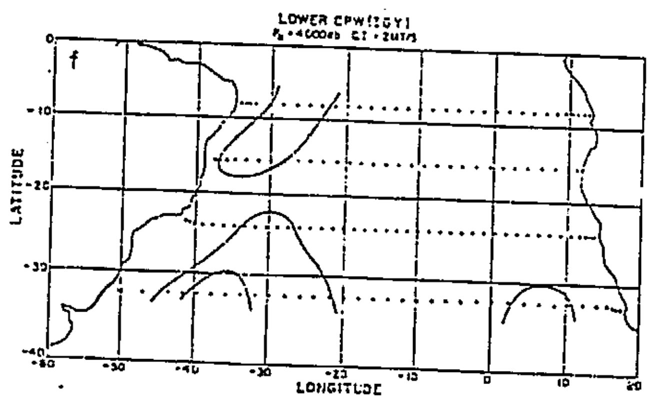

FIG. 9f. As in Fig. 9a except for the Lower CPW.

Fig. 7. Continued.

Reid's map may be compared with Fu's (1981) which shows transport streamlines for the Upper North Atlantic Deep Water, at depths from 2000 to 2250 m, close to the 2000-db surface. Each shows a westward flow between 8°S and 16°S, an eastward flow between 16°S and 24°S and a westward flow between 24°S and 32°S. Reid's map added other data of high quality that were at hand, but there were so few that the meridional gradients were not resolved. The IGY lines, spaced at 8°-latitude intervals, show alternating extrema in the streamfunction, as do the other METEOR lines. They are thus classical examples of an aliased data set, inadequate for a unique interpretation of the circulation, as the figures make clear. This is but one of many possible examples of regions in which the existent density data are inadequate to resolve the general circulation features.

Our bank of deep density data has been accumulated throughout this century, though admitteuly much more rapidly during the past 30 years. This leads to two problems when we attempt to combine these data. First, the data are of varying quality. But with effort, a careful analyst may be able to cull out the measurements having quality below what he considers adequate for the task at hand. (It should be noted that this may involve more subjective skill than objective technique.) Second, the ocean density field varies in time. Many workers have identified the effects of different short and mesoscale phenomena on density observations.

Except for restricted (limited) regions which have been extensively sampled we have little or no hydrographic information on seasonal or on year-to-year variability. It was not until 1981 that an attempt was made to resample basin-scale sections and thus estimate long-period, large scale variability in density and property distributions. Roemmich and Wunsch (1982 and submitted to Science) reported significant differences between the temperature, salinity and density distributions on zonal transects near 36° and 24°N across the North Atlantic occupied in 1957 and 1981. Swift (submitted) reports a substantial shift in the deep water temperature-salinity relation north of 50°N in the Atlantic, with the water colder by about 0.15°C and less saline by about 0.02. Variability on such large time and space scales raises doubts regarding the validity of combining separated data sets widely separated in time when representing basin scale density distributions. Moreover, as pointed out by Stommel and Schott (1977), long-period, large-scale variability in the density fields leads to unstable results when the beta-spiral (or inverse) method is applied in an attempt to estimate absolute flow fields.

4. What can a density program do for WOCE?

First, our knowledge of the general global circulation is derived principally from historical density data; thus, that information must be used as the basis of WOCE planning. To the extent that this data base, and thereby our concept of general circulation features, can be improved, the basis for such planning should be improved.

Second, use of the historical density field is the only way to relate to the interior ocean the improved pictures of mean surface circulation and forcing which are anticipated. Let us assume that WOCE could yield error-free information regarding the surface elevation, surface winds, sea-air transfers, and the understanding necessary to translate these data correctly into global fields of surface currents (or variability thereof). Even so, in general we would require knowledge of the density fields to extrapolate our knowledge of surface circulation to the ocean interior, whether this be done by numerical model or by application of the dynamic

method under the assumption of quasi-geostrophic flow.

Third, rapid, repeated density sampling could provide details in limited regions of special interest during the Experiment. Some such regions could be identified now; others will likely be identified on the basis of the new satellite measurements.

Fourth, repeated long density sections could provide estimates of longer period (seasonal or interannual) variability of selected large scale features. Such information would help put the WOCE fields in historical perspective.

Fifth, we may be able to design experiments to measure the absolute flow regime, using a combination of density, current meters and indirect current measuring devices, which would allow us to compare these measured surface currents with those estimated from satellite-dervied measures.

5. How could WOCE aid our interpretation of the density field?

There are regions of the ocean in which the deep flows are quite large. Referencing isobaric slopes to deep isobars in such regions will lead to large errors in the upper flow patterns -- in some cases reversals in the sense of flow.

One area where the deep flow results in larger sea surface slopes than obtained by relative dynamics is Drake Passage. During 1975 and 1979 absolute flow estimates were made using a combination of direct current measurements and density sections on four occasions (Nowlin et al., 1977; Whitworth et al., 1982). The across-passage height differences adjusted to direct measurements resulted in an average surface height difference of 14 cm greater than that calculated from the slope of the sea surface relative to 3000 db.

Other regions of strong deep flows are those with deep western boundary currents. We have various examples of direct measurements of speed though we don't know just how wide the deep flows are. Nevertheless, we can estimate the slope.

Arnold Gordon (1975) has measured the velocity of 5100-m depth at 56°S, 170°E, just south of New Zealand, where the abyssal antarctic water is turning northward into the Pacific. The speed of 26 cm/sec, if it extended over a width of 100 km, would correspond to a change in height of 31 cm. And, in the Atlantic, on the north side of the plateau extending eastward from South America at about 50°S, Reid's measurement of 16 cm/sec at 4933 m would give an increment of 23 cm over 100 km distance.

If we had enough data of this sort, averaged over a long enough time period, we might be able to calculate more realistic slopes and get better estimates of speed and transport. And, if we had such data to allow us to integrate these slopes into the higher latitudes where a given speed is associated with a steeper slope, we might expect to pick up at least a half meter more than we see in the relative maps.

If, on the other hand, the slope of the sea surface with respect to a level surface could be obtained in some other way, we might learn a great deal more about ocean circulation. By combining such data with the density field, we could calculate the slopes of the deep isobars, and learn more, in particular, about the abyssal circulation.

However, the degree of accuracy in the surface elevation measurements that is required to deal with the deeper flow may not be achieved in many areas. Referring to the map of steric height 3000/4000 db for the Pacific (Fig. 6) we see a range of about one centimeter in the north. At 2000/3000 db, the total range in the north (not shown) is only about 2.5 cm. In the North Atlantic and North Pacific the 2000/3500 db range is about 5 cm. While the actual ranges may differ, these maps represent the baroclinic signal. Inputs of actual gradients measured by other means (satellite altimetry, current measurements, or any other) must be equivalent in accuracy to substantially less than one cm if the deeper fields in such areas are to be considered.

Besides aiding in our interpretation of relative dynamics, WOCE might provide support and motivation for the systematic improvement of our bank of data on the large scale distributions of density and chemical parameters. Although the need for such work seems clear, the interest by the community in improving global density fields does not seem to have been high in the last decade.

6. A Density Program for WOCE

a. Improving the global density field. It seems possible that by making hydrographic and chemical measurements at a judiciously selected set of ocean stations during the next 5 or so years we could significantly improve our concept of the general circulation (and distribution of properties). Such sampling may alter substantially many of our general concepts, as well as refine our pictures and add new detail. This improved picture should be used as the basis for further detailed planning of a global ocean circulation experiment.

Even with this effort during the pre-WOCE period it is unlikely that all of the major gaps in modern global hydrography and chemical measurements could be filled. Thus, this part of the density program should continue through the experiment period.

Additional stations must include vertical sampling from the surface to near the bottom, of T, S (with modern salinometer), oxygen and nutrients at least. Sampling for other variables (TTO-type measurements) should be included as appropriate and where feasible, considering time constraints. Bottle data should be taken at intervals generally not to exceed 250 m but adequate to define the major features in water-mass and density structure.

Depending on their interests, different investigators will use different approaches in recommending where additional sampling by modern stations should be undertaken. The regions shown in Fig. 8 were selected by Arnold Mantyla (personal communication) as those most deficient in good quantity, full water column data, including complete nutrient measurements.

Joseph Reid and Worth Nowlin have suggested that many of the data sparce regions could most effectively be filled by sampling along a series of long lines. The sections that are suggested (Fig. 9) are more numerous in the Southern than in the Northern Hemisphere as there are fewer data there. Those we have drawn are meridional or nearly so. This is not because we believe such orientation is everywhere better.

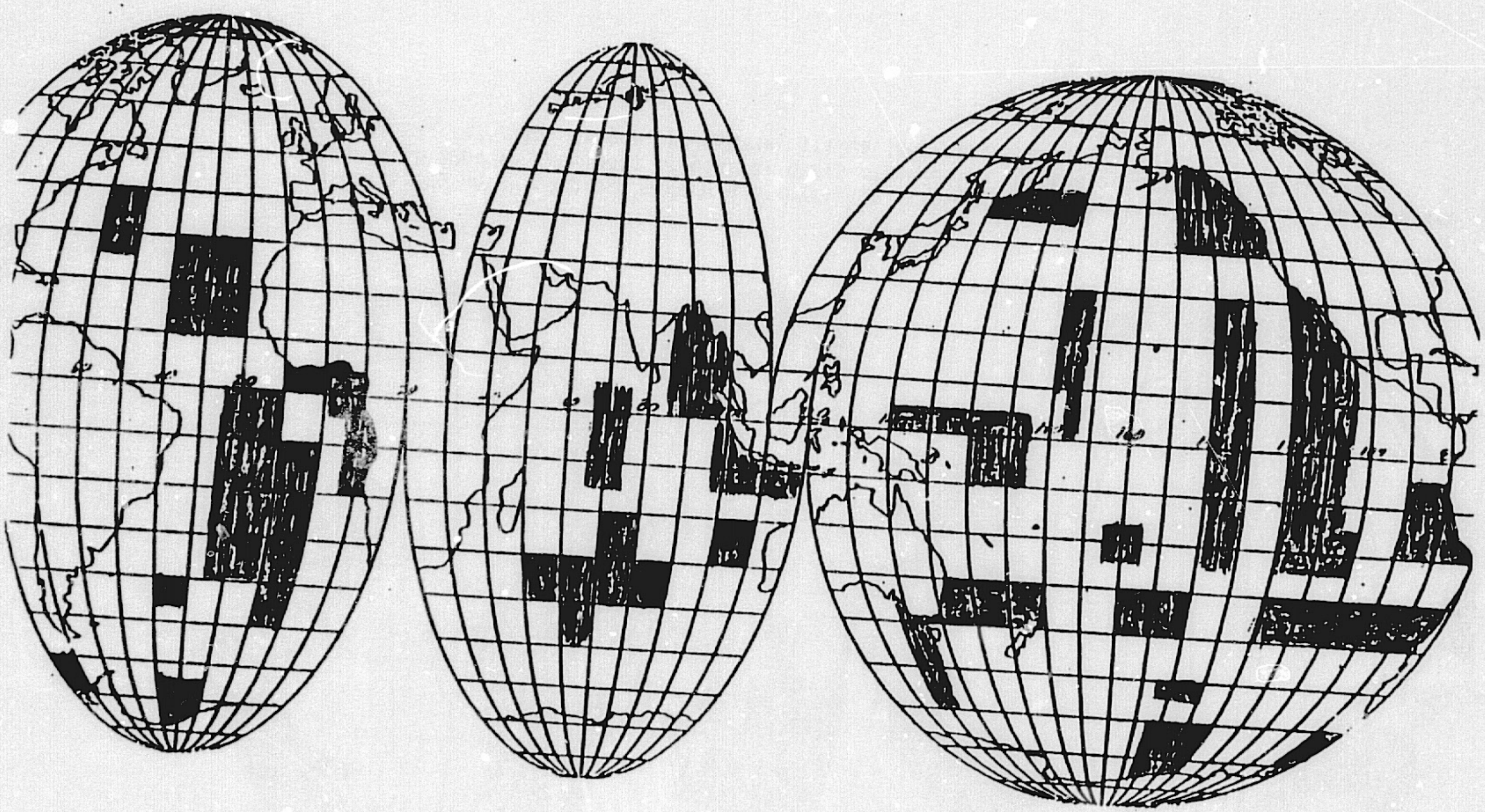

Fig. 8. Regions lacking full water column sampling including nutrients (Arnold Mantyla, personal communication).

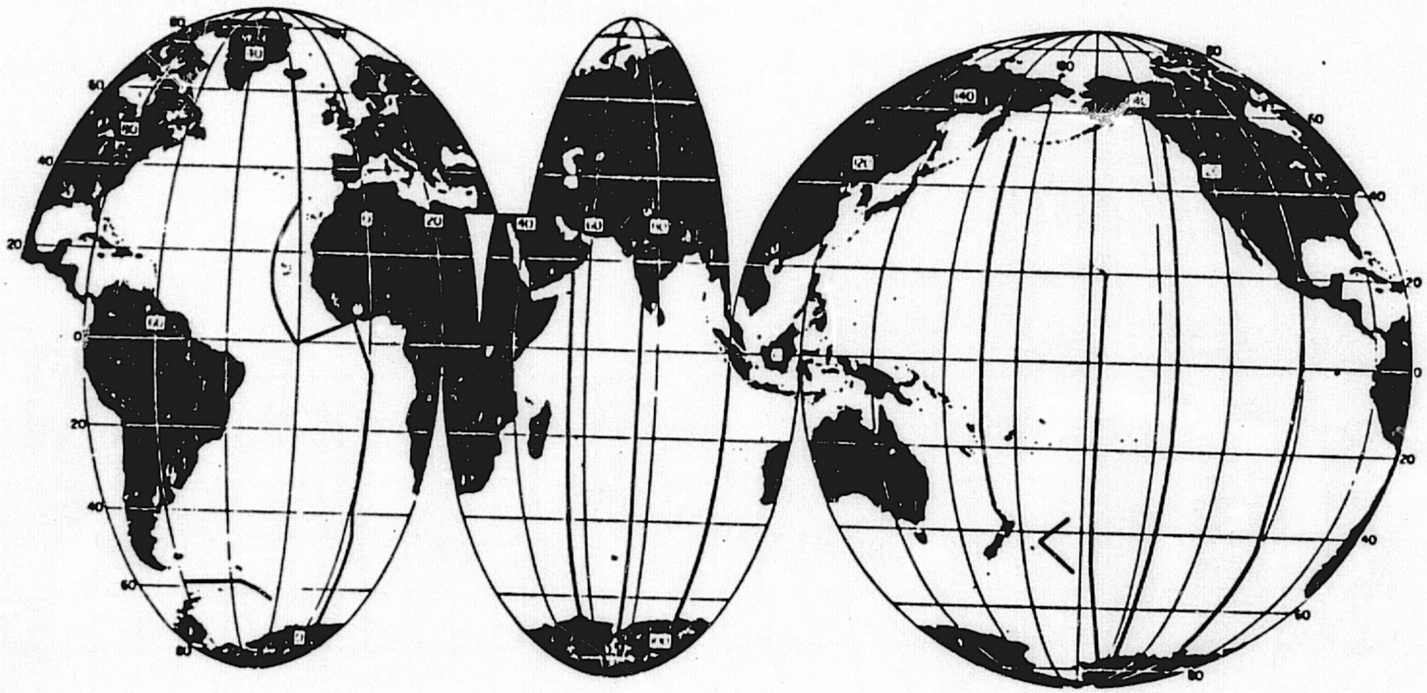

Fig. 9. Sections along which complete modern oceanographic stations at approximately 60 n. mi. spacing are suggested to improve knowledge of general ocean circulation and characteristic distributions (temperature, salinity, oxygen, phosphate, silica, nitrate, and stability).

The orientation of new sections should be determined principally by what we conjecture, or wish to learn about the flow field: they should be normal to the flow; near the continental boundaries sections should be normal to the isobaths; in the interior they should be nearly meridional, to handle the zonal part of the major gyres; and the deepest flow must be studied through sections oriented across the channels. No single section is likely to meet all these needs.

We believe the greatest limitation is in our information about the circulation of the major mid-ocean gyres, whose zonal components are not well represented by the present data bank. Recommended are those sections that we believe will give the most useful improvement on our concepts of the general circulation, and this, in turn, will help in planning other work.

We believe that in most cases horizontal spacing of stations should be at one degree of latitude, except in regions known to have narrow but significant north-south density gradients in which supplemental casts should be made. Many of these sections connect continental boundaries. Others could be extended in order to allow estimation, by the inverse method, of absolute flow fields and transport through these sections.

b. Repeated long sections. Although the recent emphasis in physical oceanography has been on oceanic variability, we have no data, except for that examined by Roemmich and Wunsch (1982) in the new zonal sections across the South Atlantic, with which to compare large-scale hydrography made at different times. In fact, no post-IGY trans-Atlantic sections were made prior to 1981. Only a single zonal section has been made across the North Pacific (Kenyon, 1983), and there are only two (the SCORPIO sections) across the South Pacific. There are no meridional sections in the North Atlantic east of 50°W, and there are none at all in the South Atlantic, although sections are planned along the Greenwich Meridian and 45°W during 1983 and 1984. No trans-Pacific meridional sections seem to exist at present, though some are planned and partial sections do exist. For all of the sections made in the past as well as those presently planned, there are basic questions regarding their representativeness of the season in which they are made or as interannual means.

During WOCE, an attempt should be made to sample some trans-oceanic sections repeatedly so as to obtain preliminary estimates of seasonality and interannual variability of the large-scale density fields.

To obtain measures of such variability it might be sufficient to establish a series of deep ocean stations at which regular hydrographic stations are repeated over long periods of time. The regularly-sampled series of Panulirus stations off Bermuda is an example that such time series can be used to obtain estimates of seasonal and longer period varaiability (Schroeder and Stommel, 1969).

c. Rapid, repeated density surveys. If we wish to determine the time-dependent flow within the ocean using currents estimated from satellite-derived measures, we need time-dependent information regarding the internal flow field. Of course, one can imagine that in some regions for limited durations the internal flow field might be measured directly. And it is possible that a measure of the variability of the internal density field may be estimated directly by remote devices, such as inverted echo sounders, or acoustic tomography. Nevertheless, considering the number of dynamically active regions which one would like to explore in the oceans, the protracted duration of a truly global ocean circulation experiment and the need for density fiel when attempting numerical assimilations of data into models, it seems that WOCE shoulu include the capability of making rapid and repeated density measurements in selected regions. The principal time-dependent role of density measurements during WOCE may

be local investigations of regions which (from other measurements) show significant differences from what we see in our (imperfect) pictures of general flow.

These measurements need not be carried out with CTD systems. That would increase the costs, limit the number of ships and organizations able to contribute and provide data which is oversampled in the vertical relative to the need. Needed for repeated realizations of the density field are quality measurements at equivalent hydrocast spacing in the vertical.

Technological developments which should be encouraged are (1) those which would allow more rapid sampling of density than is now possible and (2) those which would allow sampling from vessels with minimal capability and by personnel having only modest training. It must be kept in mind, however, that such developments will be steps backward unless they produce T, S and density measures which are of quality equal to modern hydrographic stations.

At present we have few vessels with the capability to handle reliably top-to-bottom CTD casts in all ocean depths. We have even fewer technical groups with the expertise to collect a complete suite of modern oceanographic station data (CTD profiles with rosette water samples for salinity, oxygen, nutrients, etc.). Our present level of capability, or more, will be needed to fill gaps in the global coverage by complete oceanographic stations. Only by increasing the number of vessels and groups capable of quality density work will it be possible also to undertake repeated sampling for time changes. It seems impractical to equip many vessels and train groups to make complete modern stations, if it is feasible to develop technology to carry out rapid hydrography with minimal requirements.

The "smart CTD" proposed for development by a team from Neil Brown Instrument Systems, WHOI and MIT is an internally-recording CTD, which is to be reprogrammable for various sampling schemes and to have good precision, data retrieval via connectors to a small laboratory computer and externally rechargable batteries. It is meant to be replaced, in the event of problems, rather than repaired at sea. Conceptually such a system provides the capability of measuring density from most vessels having the equipment to lower and retrieve the package -- and with only minimal technical expertise. A parallel development by this group aims toward a free-fall vehicle in which the smart CTD could be deployed. Designs call for an instrument which could make a round trip to 5000 m in one-half hour, which could collect small water samples for standardization and which could home on the surface vessel. Such developments could drastically reduce the time required for repeated trans-ocean sections as well as for regional density surveys and should be encouraged. However, their use would of course be limited to situations in which chemical measurements were not desired.

d. Sub-experiments to measure absolute flow. As part of WOCE, we may wish to select a region or section in which the absolute flow field is estimated, by a combination of direct and indirect techniques, perhaps using density and current measurements as was done for the Drake Passage (Whitworth, submitted to J.P.O.). This would allow a comparison with surface currents estimated from remote (satellite) measurements.

The region selected should be one in which the variation of surface to deep dynamic topography has a large, long-term spatial gradient and in which significant variations are expected over mesoscale time and space scales. A location fitting these criteria is a section extending from within a southern hemisphere subtropical

gyre southward across the Circumpolar Current. As we have seen the large-scale variation of sea surface topography would be great, and the region is known to be rich in mesoscale features.

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July 15, 1983

Background Paper

Satellite Wind Observations

for

National Academy of Sciences

Workshop on

Global Understanding

of the

General Circulation of the Oceans

by

James J. O'Brien
The Florida State University
Tallahassee, Florida 32306

Introduction

Since the days of early man, we have been awed by the influence of the wind on the ocean. Man has fought the wind over the sea; Man has been frightened and killed by the wind on the sea; Man has been fascinated by the effect of the wind on the sea.

When the sea is flat and calm, it is a beautiful sight; when the sea is foaming and boiling, it frightens the observer.

The wind on the sea effects every facet of the ocean's motion and, in particular, the wind and its oceanic consequences effects every activity of man related to the sea. The wind creates waves on the sea; the wind drives the ocean currents; the wind creates cold upwelled waters; the wind produces downward cascading bottom water; the wind mixes the heat of the sun from the surface layers to deeper regions; the wind moves icebergs to their demise in warmer water.

The wind is not responsibile for all ocean movement. The sun pumps more energy into tropical regions than high-latitude areas. This excess heat creates circulation patterns which redistributes the heat poleward. However, it is highly possible that this is accomplished by altering the atmospheric wind patterns and, thereby letting the wind create the ocean circulation that is required to redistribute the excess tropical heat (O'Brien, J. J., A. J. Busalacchi and J. Kindle, 1980: Ocean Models of El Nino, Chapter 7 in Resource Management and Environmental Uncertainty, ed. M. H. Glantz, (Wiley & Sons, Inc.) pp. 159-212.)

Man has always tried to measure the wind over the sea in order to anticipate its effect on oceanic activities. We rely on measurements from continental and island stations as well as ship board measurements. All of us planning to go to sea inquire routinely of the wind and wave forecast.

We weigh the advice of the weather forecast and our desire to leave the shore. When at sea, we remain aware of the sky and the sea in order to have an early warning of changing weather.

In sailing days, a good captain could read the sea and the sky and forecast the shift in the currents and the waves due to the wind. In tropical seas, he would forecast the presence of a hurricane hundreds of miles away by the strength and direction of the swell. Most wind measurements recorded in our data files are Beaufort wind estimates; these are given in Appendix A.

Man has never been capable of mesuring the wind over an entire ocean on a regular basis. The number of ships at sea, islands and shore based stations have always been inadequate for the task. The meteorologists forecast the winds on the sea but these are not accurate due to poor starting data over the oceans for their models.

It appears technically feasible to be able to measure surface winds from a satellite. In the summer of 1978, NASA launched Seasat with several experimental instruments. Almost 100 days later after 1,502 orbits Seasat died. One of the instruments on board was a scatterometer. This radar instrument is capable of deducing estimates of the wind speed and direction in a ocean patch about 50 kilometers on a side. We are not quite sure how the instrument works. It sends a radar beam to the ocean and measures the amount of back-scattered radar energy. It is supposed to be measuring the amount of waves in the ocean patch. Capillary waves are the very small 5-10 cm waves which ride on the back of the bigger waves. The premise is that the intensity, strength and quantity of these little waves are related to the wind speed at the same place and time. Some of the data from Seasat has been carefully examined by oceanographers and meteorologists and there is cautious agreement that the scatterometer is capable of

measuring wind speed in many weather situations to within 2 meters per second (plus or minus 5 miles per hour). The wind scatterometer is also capable of measuring wind direction except it almost always has a 180° bias; i.e., it can't tell North from South or East from West. Most meteorologists can remove this bias from the data by looking at a standard satellite view of the earth and noting the position of the storm centers;we need a few measurements of direction to unravel the rest.

The workshop organizers asked for comments on some specific questions.

Specific Questions:

1) With the technologies already demonstrated (Seasat), what is the demonstrated accuracy and precision of vector stress and can you estimate the accuracy of computation and put it into some context.

The Seasat program demonstrated (Brown, 1982; Jones, et al. 1982) that accuracies of $\pm 2\ ms^{-1}$ or 10% in wind speed and ±20 degrees in wind direction were achievable. A discussion of the Seasat wind parameterization algorithm can be found in Wentz (1978), Jones, et al (1978) and Schroeder et al (1982).

The NROSS scatterometer will have similar accuracies for wind speed and direction. Since it is very difficult to measure wind stress at sea, it is not planned to interpret the backscatter as wind stress even though this is highly desirable. Several viewgraphs (Appendix B) show the proposed coverage and performance specifications.

Since wind stress is at least a quadratic function of wind speed, then a 10 percent error in wind speed will mean a 20 percent error or greater in wind stress. If there are no systematic bias in the measurements, we estimate that a monthly average one degree x one degree square would have at least 16 measurements which would reduce the error to 5 percent.

It is the opinion of the S^3 committee that this global coverage will produce wind stress fields of an extremely useful nature for ocean circulation studies.

2) For realistic designs of future scatterometers what are the answers to the same questions? What sort of time/space coverage could one realistically expect?

NROSS is a single polar orbiting satellite. It, therefore, has a very poor sampling program in time and space. How bad we do not know! This is because we do not know the frequency-wavenumber spectrum of the surface wind stress field. A series of pictures in Appendix C shows the typical coverage for selected time periods.

3) What is your judgment as to what will actually be flown and when it might be flown? Assuming that such observations are found critical to WOCE are there specific steps required either to obtain upgrades necessary for scientific use, or to help fly such missions?

We expect NROSS to be launched in November 1988. No money is being allocated by JPL to learn how to build a good or better model function. It is not clear what steps the WOCE community should take to assist. In the author's opinion, the NROSS scatterometer will be more than adequate for WOCE but not for other physical oceanographic experiments.

4) What do you see as the critical issues (e.g., Does work need to be done on understanding appropriate algorithms, or are there trade-off questions about space/time coverage, ... etc.).

The critical issue is "does a scatterometer measure wind or wind stress?" No resources are allocated to answer this issue. We believe that the direction ambiguity problem which plagued Seasat has been solved; i.e., unambiguous wind vectors can be estimated for wind speeds greater than 4 meters and less than 30 m/sec. At very high wind speeds, there is no reason to expect the scatterometer to work at all for many reasons.

5) Any comments you might want to make about how such observations fit into the concept of a world circulation experiment.

If we can convince NASA to fly the S^3 scatterometer and provide resources to analyze the data into gridded weather maps, we will make a giant leap forward in estimating the main forcing function for the ocean general circulation. Considerable discussions of all the issues is contained in the S^3 report which is available.

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APPENDIX A

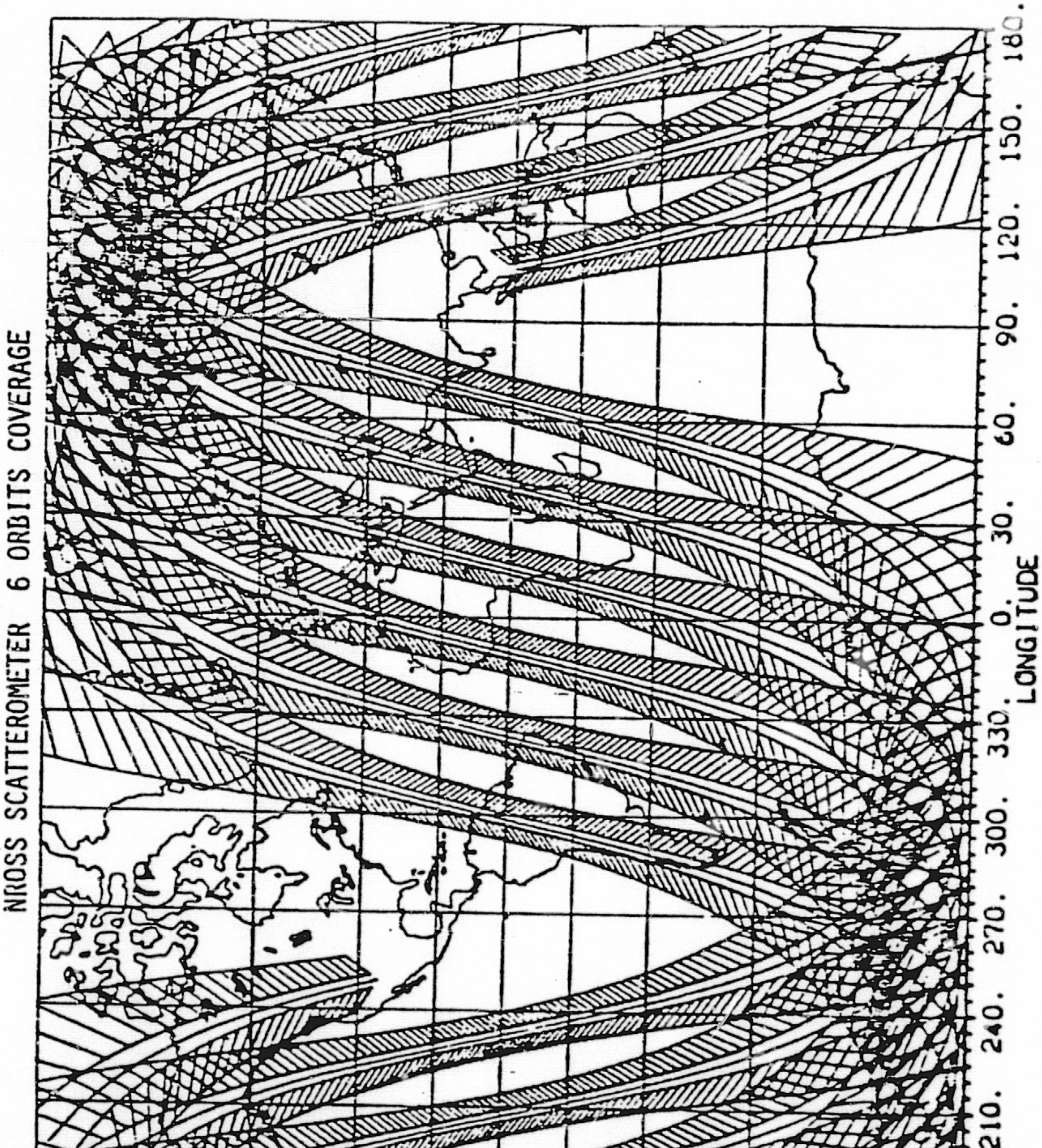
BEAUFORT WIND SCALES AND SEA DESCRIPTIONS

Beaufort Scale	Seaman's description of wind	Wind velocity knots	Estimating wind velocities on sea	International scale sea description and wave heights	International code for state of sea
0	Calm	Less than 1 knot	Calm; sea like a mirror.	Calm glassy	0
1	Light air	1 to 3 knots	Light air; ripples-no foam crests	0	
2	Light breeze	4 to 6 knots	Light breeze; small wavelets, crests have glassy appearance and do not break	Rippled 0 to 1 foot	1
3	Gentle breeze	7 to 10 knots	Gentle breeze; large wavelets, crests begin to break. Scattered whitecaps.	smooth 1 to 2 feet	2
4	Moderate breeze	11 to 16 knots	Moderate breeze; small waves becoming longer, Frequent whitecaps.	Slight 2 to 4 feet	3
5	Fresh breeze	17 to 21 knots	Fresh breeze; moderate waves taking a more pronounced long form; mainly whitecaps, some spray.	Moderate 4 to 8 feet	4
6	Strong breeze	22 to 27 knots	Strong breeze; large waves begin to form extensive whitecaps everywhere, some spray.	Rough 8 to 13 feet	5
7	High wind (Moderate gale)	28 to 33	Moderate gale; sea heaps up and white foam from breaking waves begins to be blown in streaks along the direction of the wind.		6
8	Gale (Fresh gale)	34 to 40	Fresh gale; moderately high waves of greater length; edges of crests break into spindrift. The foam is blown in well-marked streaks along the direction of the wind.	very rough 13 to 20 feet	
9	Strong gale	41 to 47 knots	Strong gale; high waves, dense streaks of foam along the direction of the wind. Spray may affect visibility. Sea begins to roll.		

APPENDIX A

Beaufort Scale	Seaman's description of wind	Wind velocity knots	Estimating wind velocities on sea	International scale sea description and wave heights	International code for state of sea
10	Whole gale	48 to 55 knots	Whole gale; very high waves. The surface of the sea takes on a white appearance. The rolling of sea becomes heavy and shocklike. Visibility affected.	High 20 to 30 feet	7
11	Storm	56 to 63	Storm; exceptionally high waves. Small and medium-sized ships are lost to view long periods.	Very high 30 to 45 feet	8
12	Hurricane	64 and above	Hurricane; the air is filled with foam and spray. Sea completely white with driving spray; visibility very seriously affected.	Phenomenal over 45 feet	9

APPENDIX B


NROSS SCATTEROMETER 6 ORBITS COVERAGE
LATITUDE
80.
60.
40.
20.
0.
-20.
-40.
-60.
-80.
LONGITUDE
180.
210.
240.
270.
300.
330.
0.
30.
60.
90.
120.
150.
180.

APPENDIX B

NROSS SCATTEROMETER
S^3/NAVY MISSION REQUIREMENTS

	S^3/NAVY	SYSTEM DESIGN COMMENTS
• WIND SPEED ACCURACY	FOR $3 \leq U \leq 30$ M/S, GREATER OF ±2 M/S OR ±10% SYSTEM CAPABILITY TO 100 M/S	U IS NEUTRAL STABILITY WIND AT 19.5 M
• WIND DIRECTION ACCURACY	FOR $3 \leq U \leq 30$ M/S, 90% OF ALL VECTOR SOLUTIONS CONSIST OF 1 OR 2 AMBIGUITIES, ~180° APART ONE OF THESE AMBUIGUITIES IS WITHIN ±20° OF THE TRUE DIRECTION	PERMISSABLE ANGULAR RANGE BETWEEN 2-AMBIGUITY SOLUTIONS IS 150° TO 210°
• LOCATION ACCURACY	EARTH-GRIDDED ±50 KM ABSOLUTE ±10 KM RELATIVE	

APPENDIX B

NROSS SCATTEROMETER
S^3/NAVY MISSION REQUIREMENTS

	S^3/NAVY	SYSTEM DESIGN COMMENTS
• SPATIAL RESOLUTION	$3 \leq U \leq 6$ M/S 100 KM $U > 6$ M/S 50 KM [NAVY: $U \geq 3$ M/S 25 KM]	50 KM RESOLUTION WILL BE PROVIDED FOR $U \geq 3$ M/S [25 KM WIND CELL CAN BE CONSTRUCTED TO MEET NAVY REQUIREMENT]
• COVERAGE	COVERAGE FROM 65°N–65°S 90% OF POINTS AT EQUATOR AND 35°N OBSERVED EVERY 2 DAYS	90% OF ALL OCEANS WILL BE OBSERVED EVERY 2 DAYS
• RAIN FLAG	σ_0 CELLS IN WHICH LIQUID ATMOSPHERIC WATER IS EXPECTED TO DEGRADE WIND ACCURACY BEYOND SPECS MUST BE FLAGGED	SSM/I DATA, WHERE AVAILABLE, WILL BE USED TO DETERMINE LIQUID WATER CONTENT IN σ_0 CELL. WHEN WATER CONTENT EXCEEDS A THRESHOLD (TBD), DATA WILL BE FLAGGED

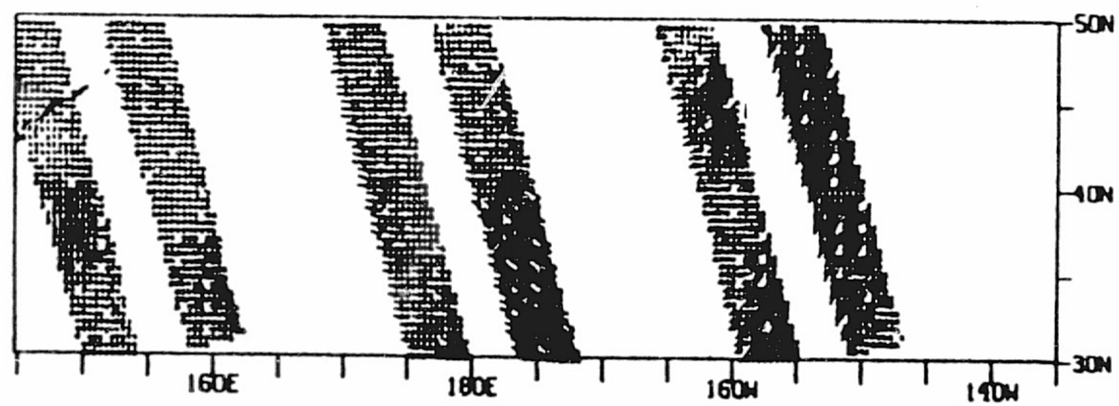
APPENDIX B

NROSS SCATTEROMETER
S^3/NAVY MISSION REQUIREMENTS

	S^3/NAVY	SYSTEM DESIGN COMMENTS
• MISSION DURATION	18 MONTHS [NAVY: 36 MONTHS]	18 MONTHS
• DATA PROCESSING AND DELIVERY	ALL OCEAN DATA PROCESSED IN 2-3 WEEKS RAW DATA ARCHIVED SYSTEM IN PLACE BEFORE LAUNCH	MULTIPLE LEVELS OF PROCESSED DATA WILL BE ARCHIVED

APPENDIX C

START AT 79 1 10 0 FOR 3.48 HOURS



APPENDIX C

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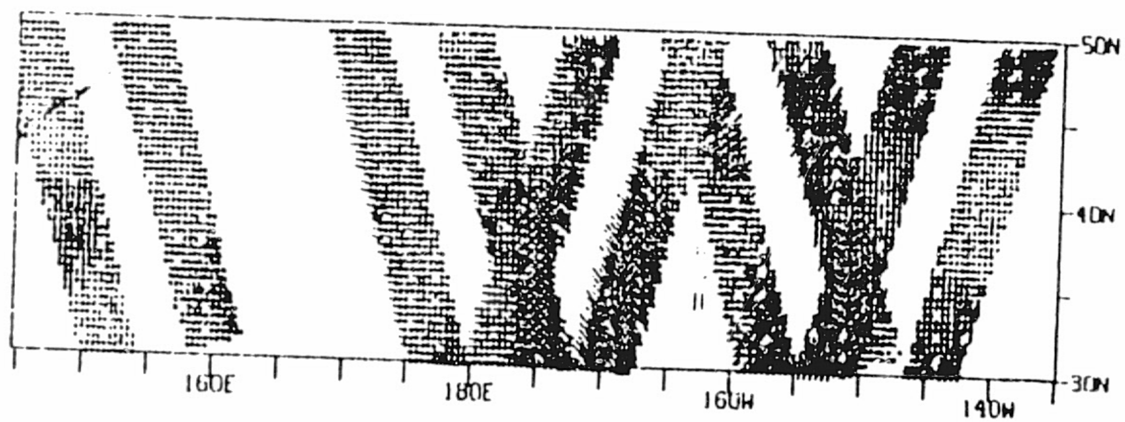

APPENDIX C

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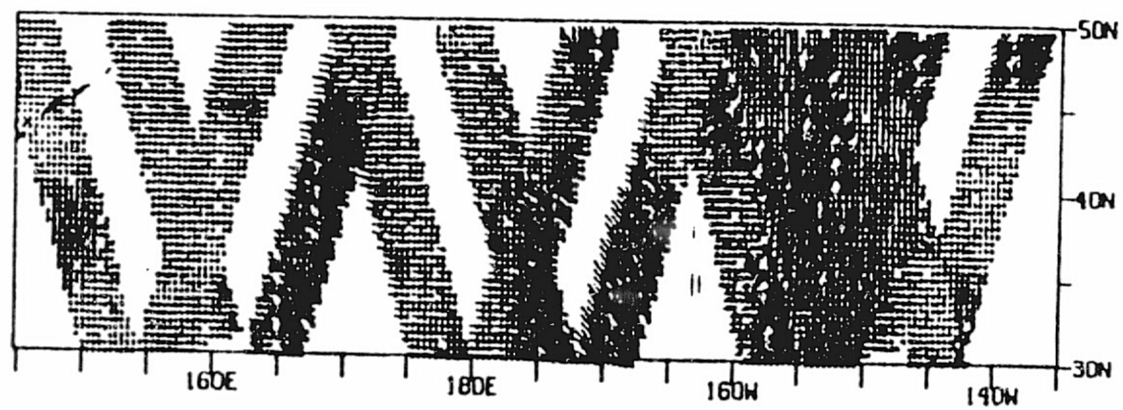
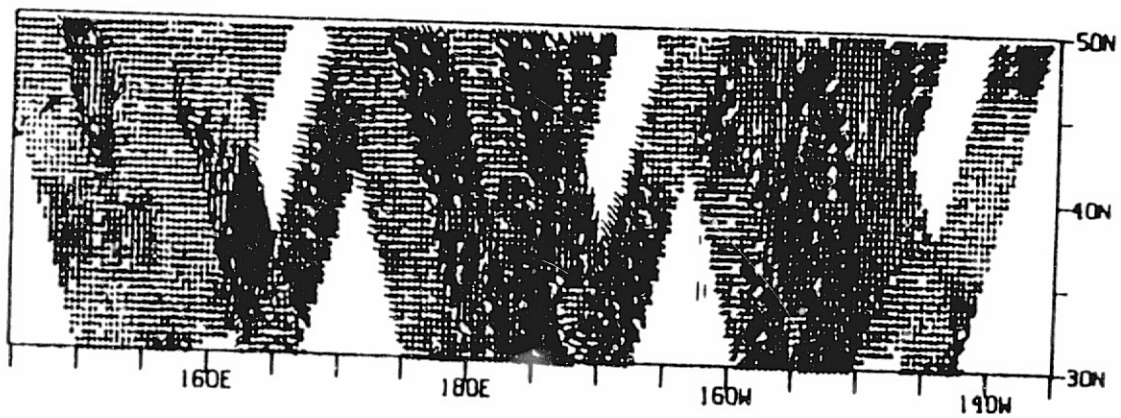


APPENDIX C

START AT 79 1 10 0 FØR 27.17 HOURS



APPENDIX C

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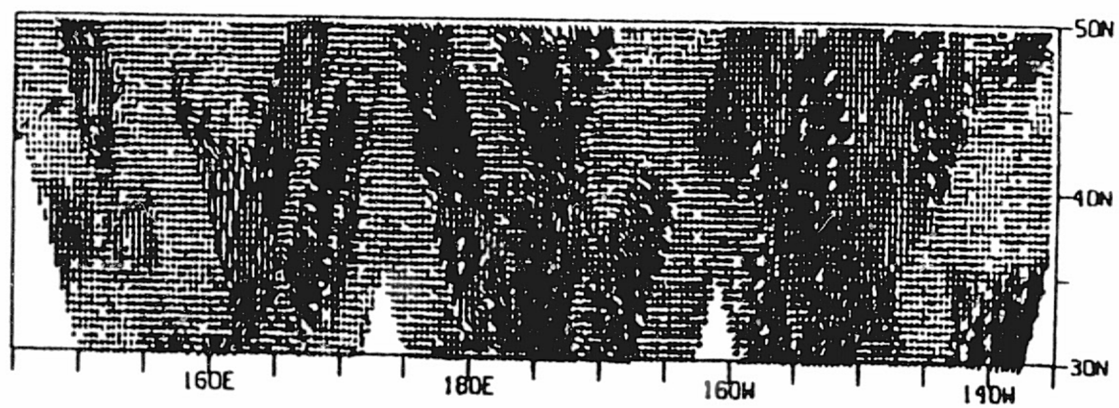
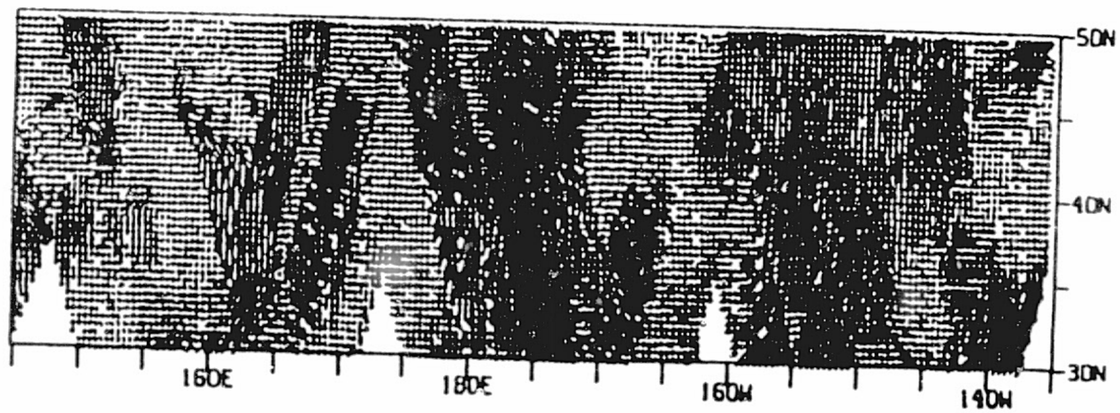


APPENDIX C

START AT 79 1 10 0 FOR 49.17 HOURS



Data Analysis and Modelling

by

Norman A. Phillips
National Meteorological Center
NWS/NOAA

A paper for presentation at the Workshop on Global Observations and Understanding of the General Circulation of the Oceans.

Woods Hole, Massachusetts

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Contents

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2. "Optimal interpolation"
3. Initialization
4. Relative usefulness of current and mass measurements
5. References

1. Background

I will discuss these questions mostly from the perspective of a meteorologist, but I will try to interpret matters for their oceanographic context. We must first recognize what at first glance appears to be a major difference. This is that for meteorologists, the primary purpose of a large-scale analysis is to form the initial data for a numerical forecast. The usefulness of a long sequence of such analyses in describing the climatology or general circulation of the atmosphere is secondary. This is because for the atmosphere, the amount of data is sufficient that direct recourse to the data is generally sufficient for these last purposes. The title of this workshop, on the other hand, appears to place first priority on the climatological and general circulation aspects of oceanic analyses. This is understandable, for obvious reasons. But I think that even the contemplated new oceanic observing systems will be an incomplete data base by themselves to answer the complicated questions contained in the phrase "general circulation of the oceans". Prediction models will be needed to interpret and extend the usefulness of the data.

For some perspective on this I recently tried to estimate the amount of data that meteorologists have. I came up with the crude estimate that for every meaningful degree of freedom in a state-of-the-art atmospheric prediction model, we get between 50 and 100 observations every 4 days on the average. The largeness of this number suggests to me that it will not be matched by any proposed oceanographic systems. I therefore think that prediction models will become every bit as important for describing the circulation of the oceans as they are in the atmosphere for making weather forecasts.

Let me begin with a simple ordering of analysis schemes, based on the way they approach different types of data.

I. Isolated

Analyse data only when and where it is observed.

II. Geographical blending with persistence (or climatology)

In regions of no data, supplement I. with an earlier analysis. (Examples: Fuglister Atlas, Mode-I, current NOAA and DOD analyses of sea surface temperature.)

III. Geographical blending with forecast.

In regions of no data supplement I. with a forecast from the previous analysis.

(Examples: Older analysis systems at NMC and other centers, previous to use of optimum interpolation techniques.)

IV. Complete blending everywhere of data with forecasts from previous analyses.

Must know statistical error properties of different kinds of data and of the forecast--the error covariances of the forecast fields and of the observations.

Options for the forecast error covariances:

A. Fixed by a gross ensemble mean. (ECMWF)

B. Fix the structure but vary the amplitude (variances) from place to place (NMC) according to the data density in the preceding analysis.

C. Predict the forecast error covariance using the model equations. (The Kalman filter approach).

Meteorological centers are making tentative steps towards seeing whether it makes sense to leave system IV.B and explore the expensive (but dangerous) promises of the Kalman approach. I believe it will be a long time however, before we are able to do this reliably. The basic techniques have been recently outlined for oceanographers by Michael Ghil

and his coworkers (Ghil, et al., 1982), so I will concentrate instead on the more elementary methods in group IV.

2. Optimum Interpolation

This technique, originally introduced to meteorology by A. Eliassen and L. Gandin, is based on a linear correction to the forecast state. Let A, F and D denote the analyzed, forecast, and data values. Let subscript k denote an analysis variable and its location, and let subscript p (or q) denote an observation variable and its location. The correction method is

$$A_k = F_k + \sum_p^{OBS} \alpha_{kp}(D_p - F_p). \tag{1}$$

The analysis weights α are of course important. Let lower case letters denote the errors:

$$a = A - A(\text{true}) \,, \quad f = F - F(\text{true}) \,, \quad d = D - D(\text{true})$$
$$A(\text{true}) = F(\text{true}) = D(\text{true}). \tag{2}$$

Equation (1) then becomes

$$a_k = f_k + \sum_p^{OBS} \alpha_{kp}(d_p - f_p) \tag{3}$$

One now asks that α_{kp} be such that the expected value of a_k^2, denoted by $\overline{a_k^2}$, be a minimum. The usual least-squares procedure gives the desired formula:

$$\alpha_{kp} = \sum_q^{OBS} \left\{ \overline{(d_p - f_p)(d_q - f_q)} \right\}^{-1} \cdot \left(\overline{f_k f_q} - \overline{f_k d_q} \right). \tag{4}$$

To calculate the weights, we must therefore know the observational error covariances $\overline{d_p d_q}$, the forecast error covariances $\overline{f_k f_q}$, and the mixed error covariances $\overline{f_k d_q}$. The most inaccessible of these is the mixed covariance, $\overline{fd}$. It is customary in meteorological practice to ignore these, although not always with justification. It is a good assumption when the observations are radiosondes, but it is less good when the observations are from remote sensors (e.g. satellite

temperatures), since these tend to have errors that are correlated with properties of the atmosphere. I believe this will be a delicate matter in the proposed oceanographic remote measurements.

Meteorologists by now have some experience with the forecast error covariance produced by their forecast models, at least in regions of good data. The practice is to collect a seasonal sample of the differences $F_q - D_q$, for 12-hour forecasts. One can then write

$$\overline{(F_q - D_q)(F_p - D_p)} = \overline{(f_q - d_q)(f_p - d_p)} \approx \overline{f_q f_p} + \overline{d_q d_p} \qquad (5)$$

For radiosondes, one can assume that $\overline{d_q d_p}$ is zero except when q and p refer to the same radiosonde ascent. Under these ideal conditions, $\overline{d_q d_p}$ can be estimated by examining the behavior of (5) as points q and p approach zero horizontal separation.[1]

If we neglect the mixed error covariance, the analysis equations (1) and (3) can be written as

$$A_k - F_k = a_k - f_k + \sum_p^{OBS} (D_p - F_p) \sum_q M_{pq}^{-1} \overline{f_k f_q} , \qquad (6)$$

by substitution of (4) for the weights α_{kp}. (M is the matrix $\overline{(d_p - f_p)(d_q - f_q)}$.)

---

[1]Results typically show (in middle latitudes) that $\overline{f_p f_q}$ is reasonably isotropic with respect to horizontal coordinates, when f_p and f_q refer to the same type of variable. The error covariances are also more-or-less geostrophic in nature. Ghil and co-authors, however, have shown from the Kalman filter approach that a noticeable non-isotropy can exist in regions where the data density changes markedly (Ghil, et al., 1981).

The important role of the first-guess error covariance matrix is that it exerts a powerful filtering role in the analysis equation (6) (Phillips, 1982). Suppose that we are able to define a set of complete orthogonal eigenvectors ϕ_{mk} that can represent all variables on the analysis grid. Any particular realization of forecast errors can then be expressed as a sum of these eigenvectors:

$$f_k = \sum_m^N \Phi_m \phi_{mk} \tag{7}$$

(This sum need not include all eigenvectors, if we wish.) The forecast error covariance matrix can then formally be expressed as

$$\overline{f_k f_q} = \sum_m^N \phi_{mq} \sum_m^N \overline{\Phi_m \Phi_m} \phi_{mk} \tag{8}$$

If we now substitute this expression into (6), we get, symbolically,

$$A_k - F_k = a_k - d_k = \sum_p (\ )_p \sum_q (\ )_{pq} \sum_m (\ )_{mq} \sum_m (\ )_{mm} \phi_{mk} \ . \tag{9}$$

This shows that the change from the forecast field to the analyzed field is a linear function of the eigenvectors ϕ_{mk} that we allowed to be included in the sum (7). The eigenvectors will differ between themselves in respect to their three-dimensional spatial scale. (7) therefore allows us to select the wavelengths we wish to include in the analysis. This property of an interpolation matrix was appreciated even in the early days of the MODE-I experiment.

An even more interesting aspect of this comes to light, however, if we are able to organize the eigenvectors on a dynamical basis. Suppose, for example, that part of the set ϕ_{mk} consists of mass and current fields that represent gravity waves, while the remaining represent essentially Rossby waves. If we only allowed the Rossby eigenvectors into the formal structure of the covariance matrix, we are assured that the resulting

changes to the forecast field caused by our analysis method will automatically possess the quasi-geostrophic and quasi-nondivergent properties of the Rossby eigenfunctions. The algebra associated with this will be more complicated in the oceans, since the eigenfunctions will not have the property of the atmospheric system--that they can be translated in longitude. The ability to guide the analysis scheme in this dynamically meaningfully way must surely be exploited, however.

These filtering attributes of the forecast error covariance of course have a reverse side: careless specification of $\overline{f_k f_g}$ may produce systematic bias in the final product.

3. Initialization

Oceanic analysis systems and forecast models with any pretext of being "global" must treat the equatorial regions. They must also treat different types of geostrophic motion--both the medium scale eddies in which the variability of the Coriolis parameter is a first-order effect, and the basin-scale motions in which the beta effect enters at zero order. This seems to require that the primitive equations must form the basis of any serious oceanic data assimilation forecast model. Purely geostrophic flow is however not an adequate initial state for a forecast model of this type, even if the bulk of the motions in the ocean are geostrophic. Use of the initialization method developed by Baer and by Machenhauer (see references in the review article by Daley, 1981) will be advisable for the ocean model just as it is for atmospheric data assimilation models. According to the rationale developed by Baer, one first considers the equations of motion written in terms of the solutions of the linearized forecast model equations. These can be dividied into the "fast" gravity modes,

with amplitudes denoted by F, and the slow Rossby modes, whose amplitudes we denote by S. The basic observation (by Machenhauer) is that the observed atmospheric field, although containing non-zero projections onto the linear fast modes, evidently is such that these fast mode components do not have a time scale as fast as their frequencies would suggest. Instead they change with time at the same slow rate as the Rossby mode components. They must therefore be "forced", so to speak, by the non-linear interactions of the Rossby modes.

If the equations of motion are written in modal form, the equation for a particular fast mode amplitude, F, would be

$$\frac{dF}{dt} = -i\omega_F F + \mathcal{P}_F \cdot NL(S,F) \tag{10}$$

The operator $\mathcal{P}_F$ is a projection of the non-linear tendencies NL onto this fast mode eigenfunction. The leading approximations in the Baer theory are

$$F^{(0)} = 0 \tag{11}$$

and

$$F^{(1)} = \frac{1}{i\omega_F} \mathcal{P}_F \cdot NL(S,0) . \tag{12}$$

This concept is proving extremely useful in current meteorological practice, because it extends to global scales the earlier ideas of initialization based on the quasi-geostrophic model of extratropical latitudes. C. Leith (1980) has provided a useful picturization of this process in his "slow-manifold" diagram (Fig. 1).

Application to the ocean will of course be greatly complicated by the effect of bathymetry and continental boundaries on the normal modes

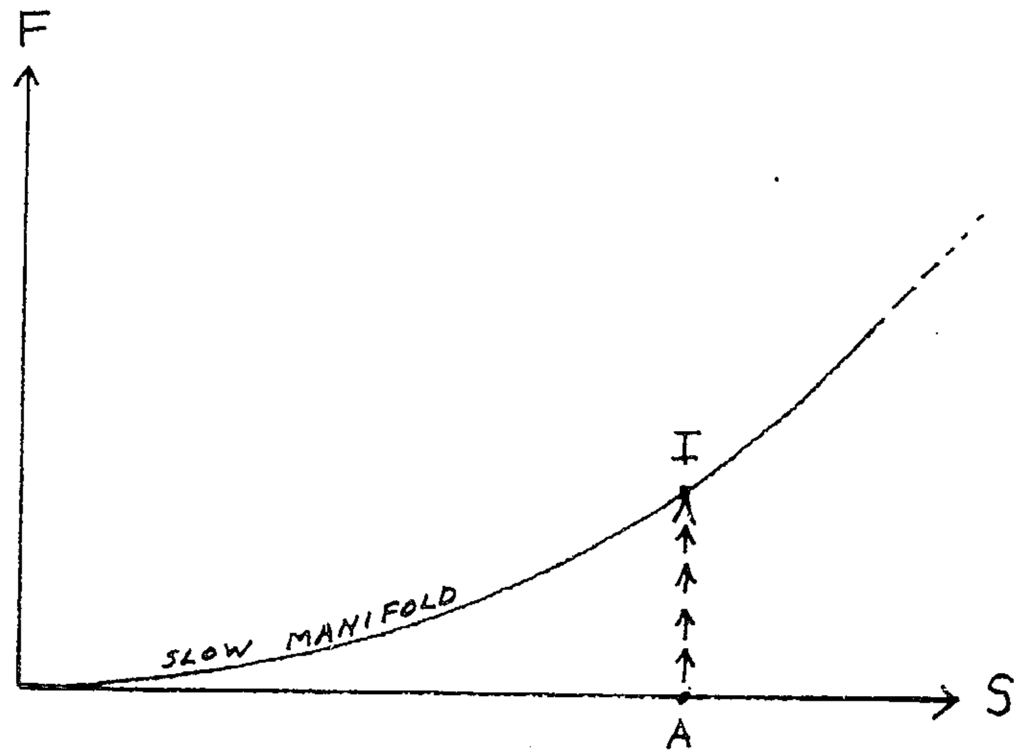

Fig. 1 – The slow manifold diagram of Leith. From an analysed field at point A containing only slow modes S, the normal mode initialization procedure of Baer derives the fast modes F to arrive at the initialized state I on the manifold. The model forecast will then remain on the slow manifold.

of the linear equations. The convenient semi-fiction for the atmosphere that the linear eigenfunctions can be translated in longitude will be unacceptable, and the eigenvectors will be more complicated. Another difference may occur with respect to the surface friction layer. In the atmosphere we seem to get by reasonably well by ignoring friction in the equations that define the eigenvectors. I'm not so sure that this will be satisfactory for the ocean, where the upper Ekman layer is a vital link in the means by which large-scale ocean motions are driven.

4. Relative Usefulness of Current and Mass Measurements

I would like to close with a simpler example of the principles of large-scale geostrophic analysis, one that might be useful in considering the relative emphasis to be placed on current measurements and on measurements of surface height or density. To put it in its most simple context, let us imagine that we have measurements of horizontal current, $\vec{v}$, and measurements of free surface height, h, for a barotropic ocean containing quasi-geostrophic eddies. Which observing system is giving us the most information? The variational analysis technique developed by Sasaki (1958) can be used to answer this.

Let $\vec{v}_o$ and h_o be the observed field at one instant. We denote the desired analyzed fields by $\vec{v}$ and h, without the subscripts. We wish to fit the data in an effective way by minimizing the area integral

$$\int w_v (\vec{v}-\vec{v}_o)^2 + w_h (h-h_o)^2 \, dA = \text{minimum} \tag{13}$$

w_v and w_h are positive weights, to be chosen most effectively. Sasaki suggested using the geostrophic relation as a constraint for this variational problem.

$$\vec{v} = (g/f)\,\hat{k} \times \nabla h \tag{14}$$

If this is introduced into (13), and we ask that h be such that (13) is a minimum with respect to any other field h, the variational calculus then yields the following differential equation for this optimum solution:

$$w_v \frac{g^2}{f^2} \nabla^2 h - w_h h = \frac{g}{f} w_v \left(\frac{\partial v_0}{\partial x} - \frac{\partial u_0}{\partial y} \right) - w_h h_0 \tag{15}$$

It is reasonable to assume that the true fields are related geostrophically, as well. Equation (15) then applies equally well to the error fields

$$\begin{aligned} \delta h &= h - h(\text{true}), \\ \delta \vec{v}_0 &= \vec{v}_0 - \vec{v}(\text{true}), \\ \delta h_0 &= h_0 - h(\text{true}). \end{aligned} \tag{16}$$

Informative solutions are obtained readily by inserting a trigonometric plan form for the variables, say, exp i ($\vec{\chi} \cdot \vec{x}$). The solution for the analysis error δh for this single realization and single wave number then has the symbolic form

$$\delta h = \delta h \left(w_v, w_h \, ; \, \vec{\chi} \, ; \, \delta h_0, \delta \vec{v}_0 \right) \tag{17}$$

We square this expression, take its horizontal average, and also take its average value over an ensemble of many realizations. If all observations have uncorrelated errors, the choices for the weights w_v and w_h that minimize $\langle \delta h^2 \rangle$ can then be determined. They are

$$w_v = \frac{1}{\langle \delta v_0^2 \rangle} = \frac{1}{\langle \delta u_0^2 \rangle} \; ; \; w_h = \frac{1}{\langle \delta h_0^2 \rangle} \, . \tag{18}$$

When these values are inserted, the final result for this wave length can be expressed as

$$\sum^{N} \frac{1}{\langle \delta h^2 \rangle} \left(= \sum^{N} \left[\frac{2 g^2 \chi^2}{f^2} \frac{1}{\langle \delta \vec{v} \rangle^2} \right] \right) = \frac{1}{\langle \delta h_0^2 \rangle} + \left[\frac{2 g^2 \chi^2}{f^2} \right] \frac{1}{(\delta \vec{v}_0)^2} \tag{19}$$

Clearly, accurate height measurements are most useful for large horizontal wave lengths, whereas accurate current measurements are most useful at short wave lengths.

Some numerical values may be of interest. The following list gives hypothetical current measurement errors that, at 40 degrees latitude, will contribute as much to analysis accuracy as height measurements that have an error of 1 cm.

L (= $\frac{2\pi}{\lvert \vec{\kappa} \rvert}$)	$\langle \delta v_0^2 \rangle^{1/2}$
10 (km)	87 (cm/sec)
50	17
100	8.7
1000	0.87

Current measurements with an rms vector <u>error</u> less than the indicated amount are more valuable than measurements of the free surface height with an error of 1 cm.

Similar analyses are of course possible for comparing measurements of vertical current shear with density measurements. An important facet, however, in this simple approach is to be careful of the actual observational error structure if the errors are correlated in space (Phillips, 1981). This will be especially necessary for remote systems, whose errors tend to have the structure of the fields themselves, instead of being random instrumental noise.

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TRACERS AND MODELING

by J.L. Sarmiento

1. Introduction

The last decade has seen a burgeoning of tracer modeling efforts and of the data available for model validation. I shall attempt to identify patterns within these developments that give indications of important directions that will be taken in the next decade. This will lead to recommendations of activities that could be undertaken within the context of a World Ocean Circulation Experiment (WOCE) to further the important goals of improving our understanding of oceanic general circulation as well as biogeochemical cycling within the oceans.

Tracer models can be classified as diagnostic or predictive. I would include in the diagnostic category any models that attempt to infer from data something about spreading or exchange rates, or biological and/or chemical processes. A simple box model which conserves tracer and has no dynamic constraints would be classified as diagnostic, as would far more complex models which do include some form of dynamic constraints. Predictive tracer models can focus on using tracers to study physical processes or to study chemical and biological processes.

The following section discusses the use of both diagnostic and predictive tracer models to study physical processes. The section after that discusses progress and future directions in the study of biogeochemical cycling. This is an area where important progress is being made at the present time, and one in

which a properly conceived WOCE could make a substantial contribution without going outside its central goals. A third section discusses tracers and future plans of tracer measurement efforts such as the Transient Tracers in the Oceans Project (TTO). A final section summarizes recommendations.

2. Tracer Modeling and Ocean Circulation

Tracer data have contributed to our understanding of ocean circulation primarily through a broad range of diagnostic techniques. Most of what we know today about the major features and time scales of oceanic general circulation derives from descriptive studies and geostrophic and box model calculations based on measurements of tracers such as salt, temperature, oxygen, and carbon-14.

The major advances that have taken place in diagnostic techniques over the past few years are due to three things:

(1) A great improvement in the quality and quantity of data available for such studies. Measurement techniques have improved greatly. A number of relatively new tracers have become accessible, partly through the improvement in measurement techniques, and partly through the introduction of new tracers into nature by bomb tests and industrial releases. The spatial and temporal resolution of high quality measurements has also improved greatly due to programs like GEOSECS and TTO.

(2) The development of new dynamic approximations, such as the Beta-spiral (Stommel and Schott, 1977), that no longer have the reference level problem of the geostrophic method.

(3) The introduction of techniques for solving underdetermined systems of equations (Wunsch, 1978). Such systems usually arise when one attempts to limit the assumptions being made in using data to infer spreading and exchange.

Much of the development of new inversion techniques has focussed on calculating the transport of properties across coast to coast sections (the techniques are, of course, not limited to such applications). One can use the transport rates thus obtained to estimate water mass conversion rates between sections, but these types of calculations are less revealing about local flow and mixing fields.

In order to obtain the local flow fields one needs to have some idea of the local tracer distribution. One way to use such data is in conjunction with local solutions to the equations of motion, such as the Beta-spiral. These solutions can be applied throughout the domain of interest. This requires a data set detailed enough to be able to calculate local derivatives of the tracer fields. Olbers et al. (1982) have used the Beta-spiral technique to estimate the North Atlantic circulation with the Levitus GFDL data set. There is no reason, in principle, why more complex dynamics cannot be used in such calculations. Indeed, Appendix 1 describes a technique we have been thinking about at Princeton that uses only tracer conservation equations to obtain a solution for the flow and mixing fields and thus avoids making any assumptions about the dynamics at all.

The further development of inversion techniques will almost certainly continue, particularly if the insatiable appetite of

such techniques for high quality data is fed by WOCE and other related studies. There are a number of important points that should be considered in deciding what types of data will be most useful:

(1) The inversions which make use of local solutions to the equations require a knowledge of the local structure of the tracer field. The zonal and meridional sections planned by WOCE will not necessarily give this information. A carefully thought out series of combined mapping and modeling exercises should make it possible to be fairly specific about the measurement resolution required in different regions of the ocean.

(2) Tracer fields vary on short time scales (order < 1 year) due to eddies, seasonal fluctuations, and other smaller scale processes. They vary on climatic time scales due to variations in the forcing conditions at the surface of the ocean. A single realization of a tracer field gives little idea of time variability, but can still be used in inverse calculations in a variety of ways. One is by ignoring long time scale changes and by objectively smoothing the short term variability. The techniques for doing this exist and can be further developed. It should be possible to use the NODC data set to get some further idea of the variability (e.g., Dantzler, 1977). I believe, however, that a few carefully chosen monitoring stations should be initiated (or continued) for measuring directly the shorter as well as the longer time scale variability over a period of many years.

(3) A list of tracers that could conceivably be measured as

part of a WOCE program is discussed below. The discussion in that section and in Appendix 1 supports my view that we should seek to obtain measurements of as many tracers as possible amongst those that can be shown to have orthogonal distributions in the oceans and a large enough signal to noise ratio. Each additional such tracer places strong constraints on inverse calculations. It is worth noting that the transient tracers are difficult to work with in inverse calculations. One requires repeated measurements to be able to estimate $\partial C/\partial t$.

The other way in which tracer data have been used in improving our knowledge of ocean circulation is as a means for validating predictive circulation models. The types of models that have been developed for predictive studies and probable directions for future developments are discussed in the technical paper by Haidvogel and Holland. Tracer studies that have been done range from abyssal circulation models using the Stommel "tour-de-force" approach, to more recent studies using primitive equation non-eddy resolving models and quasi-geostrophic, non-thermohaline, eddy-resolving models. The likely direction for future development of predictive models is eddy-resolving models with thermohaline forcing.

Tracers are proving of great importance in these studies as measures of the dispersion effected by eddies, of vertical penetration from the surface, and of long time scale horizontal spreading. Tracers help to identify what types of approximations make sense in such models as well as challenging the modelers by raising questions about newly observed features.

One of the most difficult problems that needs to be further overcome in using tracers for predictive modeling studies is an improvement of our understanding of their boundary conditions. Satellite observations will help greatly with temperature. Evaporation-precipitation remains a problem. Broecker speaks to the issue of determining boundary conditions for the biologically utilized substances in his technical report (see also below). Various efforts have been or are being undertaken by geochemists to determine appropriate boundary conditions for gases such as oxygen and carbon dioxide and for some of the anthropogenic tracers. The major contributions that WOCE can make to improving our knowledge of boundary conditions is with satellite observations of temperature, and possibly moisture content; and, if possible, by carrying out some of Broecker's suggestions for sediment trap mooring and benthic chamber studies of nutrient cycling.

3. Tracer Modeling and Biogeochemical Cycling

Until recently, models of the biologically utilized tracers have been constrained, largely through ignorance, to rather simple assumptions about the nature of biological and chemical processes. Thus, typically, we have assumed that respiration and or chemical reactions are either constant with depth or decrease exponentially. The last few years have seen an explosion of knowledge in this field due primarily to water column sediment trap research and studies of the flux of materials out of the sediments. High quality observations such as those of GEOSECS

and TTO are also playing an important role. I believe that we are on the threshold of being able to develop realistic models of biological and chemical cycling based on a real understanding of the mechanisms and time scales involved.

Such an achievement would be of great value both because of the great need we have for understanding biological and chemical processes per se, and because it would make possible a more unambiguous utilization of tracers like the nutrients for circulation studies.

One of the important areas of concern in the prediction of future climate is the level of carbon dioxide in the atmosphere. The level of atmospheric carbon dioxide as well as the oceanic oxygen content are strongly driven by the cycling of nutrients within the ocean. The prediction of future carbon dioxide levels in the atmosphere will ultimately require a detailed understanding of how nutrients are recycled within the ocean. WOCE can make an important contribution to this effort by complementing the TTO program in providing detailed measurements of water column properties, as discussed in the next section.

4. Data

Table 1 summarizes important tracers that are being measured on a fairly routine basis as part of the TTO Program. The conservative tracers are unaffected by biological and chemical processes. The starred tracers need large water volumes (~300 liters) to be measured. The half-lives of radioisotopes are given in parentheses. I will briefly discuss only tracers that

are less likely to be familiar to non-geochemists.

Radium-228 and radium-226 come from sediments and, to a limited extent, from rivers. Radium-226 is being measured primarily because it is needed for the radium-228 measurement. The data available to date shows that the signal to noise ratio of radium-226 is not great enough to justify a major measurement effort. Radium-228 originates in shelf and deep ocean sediments and, because of its short half-life, has great potential as a tracer of thermocline ventilation and deep ocean mixing and transport processes. Its input is orthogonal to that of temperature, salinity, and the transient tracers, all of which originate at the surface. Its half-life is long enough that it may be involved in a significant amount of biological cycling. This, and the fact that we do not know its source function very well, complicates its use as a tracer.

Argon-39 is produced by cosmic radiation in the atmosphere. In this respect, it is similar to natural carbon-14. Its half-life makes it of very great interest, but measurements are extremely difficult to obtain, requiring of the order of 1000 liters of water and counting times of the order of 1 month.

Krypton-85 is considered to be somewhat similar to the freons in that it is anthropogenic (it is produced by nuclear reactors) and has been increasing exponentially with time. It also requires large volumes of water to measure.

Cesium-137 was produced by bomb tests and has a similar time history as tritium. Cesium-137 is also now being released by nuclear reprocessing plants, primarily at Windscale in the United

Kingdom. This has produced a northward moving plume which hugs the European coast and is now penetrating into the region north of Iceland.

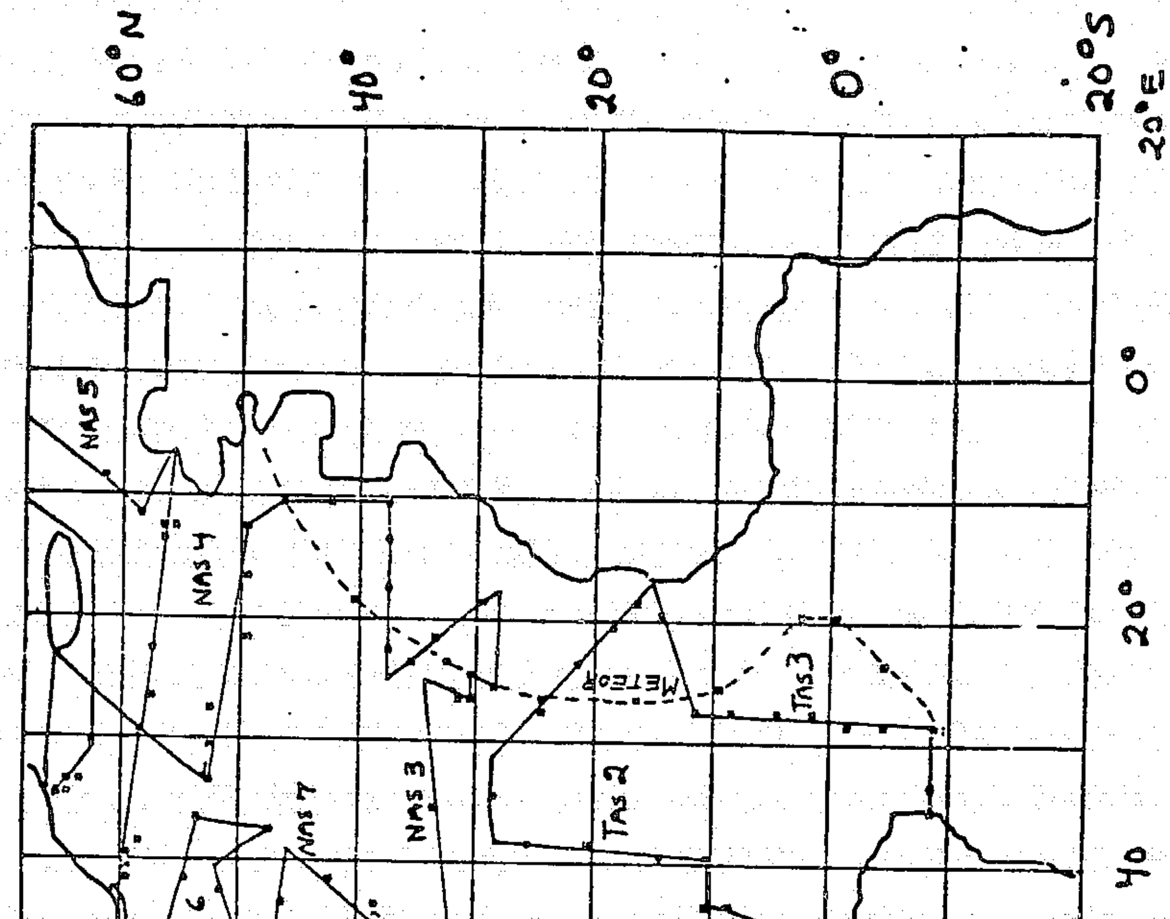
Worldwide measurements of all these tracers except Argon-39, Krypton-85, and freons, were obtained primarily along north-south sections by the GEOSECS program during the 1970's. The TTO Program is now repeating the measurements, partly to study the evolution of the anthropogenic tracers through time, and partly to provide the third (zonal) dimension that was only marginally covered by GEOSECS. Figure 1 shows the locations of stations where measurements of the TTO tracers have been obtained during the past 3 years. Figure 2 gives locations of stations planned for the South Atlantic during 1984-1986. The plans of TTO beyond the South Atlantic Study are not settled as yet. It now appears that the French will be running a TTO style program in the Indian Ocean for 3 months a year during 1984-1986. The logical next area for TTO to continue its work would thus be the Pacific Ocean.

What tracers should be measured as part of a WOCE Program? The following are important considerations:

(1) Which tracers provide useful additional information relative to the others? I shall speak to this below.

(2) What measurement resolution is desired, and to what extent is this resolution already being provided by TTO? I believe that this matter should be addressed by a modeling/mapping group that should be set up for this purpose.

(3) Ease of measurement. Argon-39 is impractical for a WOCE


60°N
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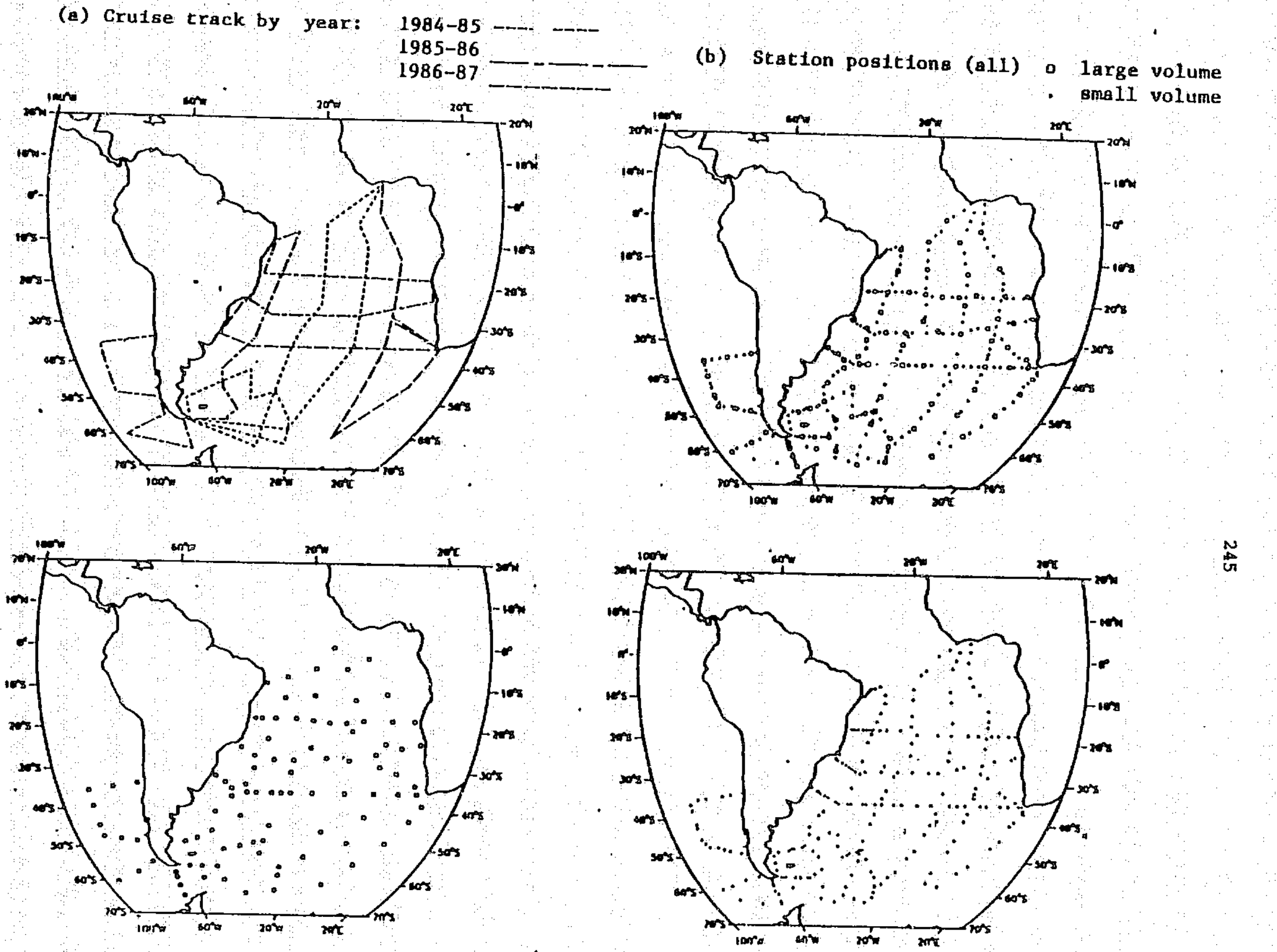
NAS 4
NAS 5
NAS 6
NAS 7
TAS 1
TAS 2
TAS 3
METEOR
TASI CRUISE

FIG. 1



(c) Large volume stations

(d) Small volume stations

FIG 2

type program, and all the tracers requiring large water volumes are difficult. It will probably be necessary for the TTO Program to continue to have major responsibility for these measurements, though a close coordination of efforts in the Pacific would probably result in considerable savings.

The primary criteria for deciding which tracers provide additional information relative to others is orthogonality of distribution and, for non-stable and/or non-conservative tracers, whether or not they either provide a time-clock or information about chemical and/or biological processes that is important to have. Appendix 1 discusses one way of using tracers that illustrates how one can use a variety of tracers to constrain what the mixing and flow fields have to be. The rigorous way to estimate the degree of orthogonality is to calculate the Jacobian of one tracer with respect to another. We plan on calculating this along with error estimates using the TTO data set in the North Atlantic. One can gain an initial impression of the orthogonality by discussing source functions and studying property-property plots. I will discuss each of the tracers in Table 1 in turn.

The boundary conditions for temperature and salinity are well known. Property-property plots of these tracers in the GEOSECS Atlases show that, within error, temperature and salinity are easily able to separate water samples taken in the upper ocean. Even in the deep ocean, the measurement precision is great enough to separate most water samples. These are the only two stable conservative tracers (ignoring geothermal heating,

which may be quite significant in the sluggish deep Pacific).

Phosphate and nitrate are linearly correlated with each other, though their distributions are quite different from that of temperature and salinity. Their surface boundary conditions are determined by biological uptake, which is relatively independent of temperature and salinity. Their distributions are nearly identical because the biological processes of fixation (at the surface) and respiration (important in the interior) act on both of them in the same way; the Redfield ratio of nitrate to phosphate in organisms is nearly constant everywhere. However, if one looks at the distribution of these two tracers in more detail, one finds many regions where there is an important signal in one tracer relative to the other. Nitrate is involved in an additional set of reactions, nitrification-denitrification, which does not affect phosphate.

Phosphate and nitrate thus can yield information about respiration rates (J_1) and denitrification rates (J_2) in the interior (see Table 1). If we are able to estimate J_1 or J_2 by techniques such as sediment traps, these two tracers will give us a measure of time. These tracers can also be used as conservative tracers by eliminating J_1 by using the apparent oxygen utilization rate (Saturation oxygen level calculated from the potential temperature minus the observed oxygen). Oxygen is also involved in respiration and is utilized by organisms in a relatively constant proportion to phosphate and nitrate. One can use the oxygen correction to calculate the nutrient content the water had when it left the surface. These values for preformed

nitrate and phosphate will not be correlated with temperature or salinity, though they will be correlated with each other to the extent that their Redfield ratio is everywhere constant, and none of the nitrate reactions are occuring.

Alkalinity and total carbon can be used in combination with the nutrients to obtain J_4, the dissolution of calcium carbonate in the water column. Total carbon and alkalinity provide information which is independent of salinity. However, the measurements are difficult to make and have low precision so that they are not particularly useful as conservative tracers in the same form as preformed phosphate or nitrate.

Silica provides information on J_5, the dissolution of silica in the water column. There is no tracer that can be used in combination with silica to give a conservative tracer.

Unless we can get an independent estimate of the J's, none of the foregoing tracers will give us a direct measure of time history. That is the role of the naturally occuring radioisotopes and the anthropogenic tracers. The most valuable of the naturally occuring radioisotopes is carbon-14. Measurement precision of radiocarbon is equivalent to a time resolution of 20 to 30 years. This is too long to be of much use in the upper ocean where, in any case, the contamination by bomb-produced radiocarbon is very great. Radium-228 has a measurement precision equivalent to a time resolution of approximately 6 months. It has great potential, but is not yet proven. Its complicated source function and possible involvement in biological cycling may make its use difficult.

The anthropogenic tracers provide extremely important resolution of a few years for exchange between the surface and interior waters. The different time histories of inputs of tracers like the freons (exponentially increasing) and tritium (a pulse) make these tracers useful in combination with each other. The tritium/helium-3 pair is unique in providing a dating technique precise to about 1 month. The interpretation of this date is ambiguous in the presence of diffusion, particularly where the parent has not yet achieved a steady distribution.

5. Summary of Recommendations

I have three major recommendations:

(1) A modeling/data mapping group should be set up to consider in detail the measurement resolution that would best serve the needs of the diagnostic and predictive tracer modelers.

(2) All the tracers shown in Table 1 would contribute to the goals of a WOCE program. There needs to be a close coordination of the TTO Program with WOCE in order to optimize resources and obtain the measurement resolution recommended by the modeling/data mapping group.

(3) The problem of determining surface ocean boundary conditions is as important for nutrient and other tracer balances, as it is for heat and water balances. There are important contributions that can be made in this area by satellite observations of physical (e.g., wind speed, which affects gas exchange) and biological (e.g., color, which may be relatable to productivity) surface ocean properties. The use of satellite data

will require an improvement of our understanding of the fundamental relationships between the measured properties and the processes we wish to quantify. This is an area which needs to be looked at more carefully, probably by a series of process oriented studies.

Table 1

Chemical Tracers of Ocean Circulation

	Conservative	Non-Conservative	Source/Sink
Stable Tracers	Temperature	Oxygen	$(O_2/PO_4)\cdot$J1 (respiration)
	Salt	Phosphate	J1
		Nitrate	(NO_3/PO_4)J1 + J2 (denitrification)
		Silicate	J3 (SiO2 dissolution)
		Total Carbon	(C/PO_4)J1 + J4 (CaCO3 dissolution)
		Alkalinity	$2\cdot$ J4
Naturally Occuring Radioisotopes	* Radium-228 (t1/2 = 6 yrs)		$(Ra/SiO_2)\cdot J_3$ (?)
	* Argon-39 (t1/2 = 270 yrs)		
		* Carbon-14 (t1/2 = 5730 yrs)	$(C\text{-}14/PO_4)\cdot$ J1 + J4
		Radium-226 (t1/2 = 1600 yrs)	$(Ra/SiO_2)\cdot$ J3 (?)
Anthropogenic Tracers	Tritium/Helium-3 (t1/2 = 12.4 yrs)		Bombs
	* Carbon-14		Bombs + $(C\text{-}14/PO_4)\cdot$ J1 + J4
	* Cesium-137 (t1/2 = 30 yrs)		Bombs and nuclear fuel reprocessing
	* Krypton-85 (t1/2 = 11 yrs)		Nuclear reactors
	Halocarbons (Freons)		Industry
	Total Carbon		Industry + $(C\text{-}14/PO_4)\cdot$ J1 + J4

* Require large volume sampling (approximately 300 liters).

Appendix 1

Calculation of Velocity and Mixing Coefficients through Simultaneous Use of Tracers

M. Kawase (Princeton University)

Below I describe a method for calculating the velocity and mixing coefficients in the ocean by simultaneously using several tracers. The technique bears similarities to that developed by Rooth and Ostlund (1972). The essence of the method lies in using some of the steady state tracer fields as spatial coordinates, and in transforming other tracer fields to this set of coordinates, thus eliminating the advection terms in the tracer conservation equations. The equations are derived in geodesic coordinates, though there is no reason why they could not also be derived in density coordinates (e.g., Rooth ~~and~~ Fine and Östlund, 1980).

Assume we have three conservative, steady state tracers α, β, γ, which are independent of each other (in the sense that their Jacobian does not vanish) in a certain region of the ocean. Assume, also, that we have non-conservative steady state tracers $C_1, C_2, C_3 \ldots$ with decay terms $J_1, J_2, J_3 \ldots$ The equations of tracer conservation are

$$\bar{u}\cdot\bar{\nabla}\alpha = \bar{\nabla}_H (K\bar{\nabla}_H \alpha) + \frac{\partial}{\partial z}\left(D\frac{\partial}{\partial z}\alpha\right) \quad (1)$$

$$\bar{u}\cdot\bar{\nabla}\beta = \bar{\nabla}_H (K\bar{\nabla}_H \beta) + \frac{\partial}{\partial z}\left(D\frac{\partial}{\partial z}\beta\right) \quad (2)$$

$$\bar{u}\cdot\bar{\nabla}\gamma = \bar{\nabla}_H (K\bar{\nabla}_H \gamma) + \frac{\partial}{\partial z}\left(D\frac{\partial}{\partial z}\gamma\right) \quad (3)$$

$$\bar{u}\cdot\bar{\nabla} C_i + J_i = \bar{\nabla}_H (K\,\bar{\nabla}_H C_i) + \frac{\partial}{\partial z}\left(D\frac{\partial}{\partial z}C_i\right) \quad (4, i)$$

We can transform the spatial coordinates (x, y, z) to (α, β, γ), since $\partial(\alpha,\beta,\gamma)/\partial(x,y,z) \neq 0$. Applying this to equations $(4, i)$ ~~above~~ and subtracting from them $\frac{\partial C_i}{\partial \alpha}\cdot(1) + \frac{\partial C_i}{\partial \beta}\cdot(2) + \frac{\partial C_i}{\partial \gamma}\cdot(3)$ gives

$$J_i = K\left(\frac{\partial^2 C_i}{\partial \alpha^2}|\nabla_H \alpha|^2 + \frac{\partial^2 C_i}{\partial \beta^2}|\nabla_H \beta|^2 + \frac{\partial^2 C_i}{\partial \gamma^2}|\nabla_H \gamma|^2\right)$$
$$+ D\left(\frac{\partial^2 C_i}{\partial \alpha^2}\left(\frac{\partial \alpha}{\partial z}\right)^2 + \frac{\partial^2 C_i}{\partial \beta^2}\left(\frac{\partial \beta}{\partial z}\right)^2 + \frac{\partial^2 C_i}{\partial \gamma^2}\left(\frac{\partial \gamma}{\partial z}\right)^2\right) \quad (5)$$

Thus, if we have two tracers whose decay terms are known (for example, radioactively decaying), we can in principle calculate local K and D. This in turn enables us to calculate the velocity components and J 's which are not known. Even if we do not know completely any of the J terms, we can obtain some information about K and D if we can place bounds on the J terms.

It is possible to include time-dependent (transient) tracers in C_i's, provided that we know the local tendency of the tracer. It simply gives an additional term $\partial C_i/\partial t$ on the LHS of equation 5. Use of time dependent tracers for the coordinate tracers is not recommended since this requires the calculation of tracer tendencies at fixed (α, β, γ) points (which will be moving in real space).

In order for this method to be succesful, we must have at least three independent tracers (or two if the problem is solved on isopycnal surfaces). Moreover, since we make use of the second

derivatives of one tracer field with respect to another, the tracer fields should be reasonably smooth.

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Rooth, C.G. and H.G. Ostlund, 1972. Penetration of tritium into the North Atlantic thermocline. Deep-Sea Res., 19, 481-492.

Chemical Tracer Measurements

R. F. Weiss
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

Introduction

Despite the best of intentions, the inclusion of a useful range of geochemical tracer studies in a World Ocean Circulation Experiment can be a massive logistic undertaking. To decide which tracers should best be included in such an experiment, one must balance potential scientific usefulness against a number of practical considerations. Does the tracer vary independently from other measured tracers, and does it vary on a spatial or temporal scale which is useful for the region of the ocean being studied? Are the complications associated with defining the source function or evaluating *in situ* chemical or biological interactions tractable? What amount of water is required for an analysis, and is there a need for special samplers or sampling procedures? Can the measurements be made at sea, and if so, what is the number of analysts and the amount of laboratory space required? What is the dynamic range of the tracer -- that is, what is the range of expected variation divided by the measurement precision? How rapidly can the measurements be made, and at what cost, considering present limitations in equipment and trained personnel? For tracers which are not at steady state, how frequently should measurements be repeated?

These problems are familiar to geochemists who have been engaged in large scale global tracer studies. Even with the

largest vessels in the academic fleet, logistic considerations have forced severe limitations in the number of tracers studied and in the density with which these could be sampled. Thus, while there is broad support for closer integration of this work with physical oceanographic studies, I do not believe there is a consensus regarding the best way to carry this out. Characteristics of some of the more useful tracer measurements are listed below, followed by some recommendations for their integration into WOCE.

<u>Salinity, Oxygen and Nutrients</u>

These chemical measurements have become a routine part of large-scale circulation studies, and their incorporation into WOCE need not be defended here. However, with the possible exception of salinity, their accurate measurement cannot be taken for granted. Routine oxygen measurements are still carried out by manual colorimetric titration with visual endpoint detection, and the accuracy and precision of the results is dependent on the skill and judgement of the operator. There is room for significant improvement here, such as by automation of the titration with colorimetric endpoint detection. There is also a potential for improvements in oxygen sampling techniques to reduce the possibility of atmospheric contamination. While nitrate, phosphate and silicate are now generally measured by automated colorimetry (Autoanalyzer), the results are still dependent on factors which are difficult to control adequately, such as temperature and reagent quality. The reliability of these measurements could be improved significantly by use of improved automatic colorimetry

equipment in which these variables are more closely controlled, or perhaps by use of recently developed techniques in automated precision ion chromatography.

Radiocarbon and the Carbonate System

Much of what geochemists have learned about large-scale ocean circulation has come from the study of radiocarbon (^{14}C) distributions. In the deep basins at low and mid latitudes ^{14}C distributions are dominated by the decay (mean life 8200 y) of cosmic ray-produced radiocarbon, and in shallow and high latitude waters ^{14}C distributions are dominated by the injection of radiocarbon produced by atmospheric testing of thermonuclear weapons. Surface water radiocarbon equilibrates with the atmosphere on a time scale of 5-10 years.

The interpretation of measured ratios of radiocarbon to stable carbon ($^{14}C/^{12}C$) is complicated by the fact that the concentrations of both forms of carbon are affected by biological degradation of particulate and dissolved organic matter, and by dissolution of calcium carbonate tests in the water column and at the sediment water interface. Thus, in addition to measuring the dissolved $^{14}C/^{12}C$ ratio, it is necessary to characterize the dissolved inorganic carbonate system (total inorganic carbon, carbonate alkalinity, pCO_2) and the oxygen and nutrient systems to interpret the data correctly . These additional measurements add significantly to the logistic requirements of studying radiocarbon distributions.

At present, the most accurate $^{14}C/^{12}C$ ratio measurements are made by shorebased gas counting techniques. The relative precision of these measurements is about 0.5%, or about 1/60th of the range found in the oceans. However, each measurement requires the collection and shipboard extraction of about 200 liters of water. This requires additional ship time for special casts with large volume samplers, as well as 2 or 3 technicians and space to carry out the shipboard extractions and carbonate system measurements.

This situation may be greatly improved by recent technological developments. Measurements of $^{14}C/^{12}C$ ratios by accelerator techniques which require only 50 ml of seawater are now approaching the precision of gas counting techniques. Also, very accurate shorebased measurements of total inorganic carbon and carbonate alkalinity on stored 1 liter seawater samples are now possible. However, these new measurements are still extremely tedious, costly, and time-consuming, and a major new development effort or expansion in scale is needed to make them suitable for large-scale oceanographic use.

<u>Tritium and Helium Isotopes</u>

Tritium (^3H) is a nearly ideal transient oceanographic tracer: it is a radioactive constituent (mean life 18 y) of the water molecule and it is almost totally man-made. As with radiocarbon, its principal source is the atmospheric testing of nuclear weapons. However, because of rapid tropospheric removal, its oceanic source function is more strongly dependent on latitude, with higher inputs at high latitudes and in the northern

hemisphere. Surface water tritium concentrations do not equilibrate with the atmosphere, but rather are controlled by vapor exchange and rainout deposition rates. The spatial and temporal distributions of the source function are moderately well known.

With conventional shorebased counting and electrolytic enrichment techniques, tritium can be measured with a relative precision of about 1% and a detection limit of about 0.03 Tritium Units (TU), where 1 TU is defined as a tritium/hydrogen ratio of 1 x 10^{-18}. High latitude northern hemisphere surface waters have a tritium concentration of about 5 TU. Recent developments in tritium measurement techniques, based on the shorebased mass spectrometric measurement of the growth of the daughter product helium-3 (^3He) in stored samples, have significantly lowered the detection limit of tritium, and concentrations as low as 0.001 TU may now be measurable. The sensitivity is dependent on the volume of the sample and the length of storage, as well as on the degree of contamination. Both the mass spectrometric and the low level counting techniques can generally be carried out on less than one liter of water.

Despite these obvious advantages, it is important to recognize that there are only a few laboratories in the world that are capable of carrying out a large scale program of accurate measurements. To undertake a higher density oceanic tritium measurement program than has already been incorporated into programs such as TTO will require a major commitment to expand our present measurement capabilities.

Measurement of oceanic distributions of ^3He, the stable and inert daughter of tritium, extends the utility of tritium measurements via the tritium - helium-3 dating concept to the study of transport and mixing processes with shorter timescales. The production of radiogenic ^3He in tritium-bearing waters must be evaluated against background ^3He concentrations resulting from atmospheric solubility equilibrium at the sea surface. Enrichment in the $^3He/^4He$ ratio thus provides a useful diagnostic of excess ^3He. Helium isotope measurements can be carried out with the same shorebased equipment as the mass spectrometric tritium measurement technique.

In the deep ocean, helium isotope measurements provide a different and unique type of tracer information. Hydrothermal fluids associated with volcanism and seafloor spreading inject large quantities of primordial ^3He into the deep sea. This process is unique in that it provides a stable and conservative tracer which is injected at locations which tend to be at mid-depth. For example, $^3He/^4He$ ratios about 50% in excess of atmospheric equilibrium have been found over the East Pacific Rise at a depth of about 2.5 km, and a plume of excess helium has been traced to the west for several thousand kilometers. The present deep-sea helium isotope measurements are sparse, and many areas of the ocean remain unexplored.

Halocarbons

The potential value of dissolved atmospheric CCl_3F (F-11) and CCl_2F_2 (F-12) as conservative time-dependent tracers of ocean

circulation and mixing has been recognized for some years. These compounds are extremely stable in natural waters, they have no natural sources, and their histories of release to the atmosphere are fairly well known. Because of their importance in stratospheric chlorine chemistry and in the modulation of the ozone layer, the global atmospheric distributions of these gases are closely monitored. Unlike the transient increases in tritium and radiocarbon which are used as tracers on a similar decadal time scale, the atmospheric distributions of F-11 and F-12 are not strongly dependent on latitude, and their surface water concentrations can be expected to come into solubility equilibrium with the atmosphere on a time scale of about a month. The presence of separate halocarbons with different rates of increase in the atmosphere provides additional constraints on time-dependent models of subsurface circulation and mixing.

There have been major improvements in oceanic halocarbon sampling and measurement techniques in recent years. Measurements are made aboard ship on samples of about 30 ml using electron capture gas chromatography, which provides a detection limit on the order of 1 part in 10^{15}. The precision of the seawater measurements is about 1%, and the present detection limit compared to polar surface water values is about 1/400 for F-11 and about 1/200 for F-12. With present technology, two shipboard analysts can measure F-11 and F-12 in about 40 sample per day (a density which approaches that of routine hydrographic sampling), and the preliminary data are available shortly after leaving a station. The major source of contamination in low level F-11 and F-12

measurements now appears to be in the Niskin sampling bottles. These effects can be reduced by using larger bottles with more favorable surface-to-volume ratios (at least 10 liters are preferred), or better still, by undertaking to construct noncontaminating bottles for routine hydrographic use. It should be emphasized that such halocarbon measurements are still in the development stages, but that with sufficient effort they could be incorporated into WOCE as routine measurements.

The remarkable sensitivity of shipboard electron capture chromatography for the measurement of halocarbons has also raised the possibility that such compounds intentionally injected into the water column could be used as purposeful tracers of ocean circulation and mixing. With present techniques, 50 kg of halocarbon dispersed in 500 km^3 of seawater could be measured with a precision of 1%.

Argon-39 and Krypton-85

These two radioactive noble gases deserve special mention for their potential value as chemically inert ocean tracers. Argon-39 (^{39}Ar) is a "non-transient" atmospheric constituent produced by cosmic ray bombardment of ^{40}Ar. Like other inert gases, its concentration equilibrates with the atmosphere on a time scale of about 1 month. It decays with a mean lifetime of 388 y, which is ideally suited for the study of thermocline penetration and deep ocean circulation and mixing. Unfortunately, the measurement of oceanic ^{39}Ar distributions with present technology is exceedingly difficult. Seawater samples of about 2000 liters are required for

each analysis, and special counters are required which now exist only in Bern, Switzerland. The maximum turnover rate is about 4 samples per counter per year, and the precision is about 5% of the modern $^{39}Ar/^{40}Ar$ ratio.

There is the potential that the sample requirements could be reduced to a few liters using laser spectroscopy, but this would require a major development program, both in purification techniques for the separation of ^{39}Ar from the large background of ^{40}Ar, and in the laser counting techniques themselves. The time scale for such improvements, even with adequate funding and personnel, may be several years.

Krypton-85 (^{85}Kr) is a transient tracer with a mean life of 15 years which is released to the atmosphere as a product of nuclear fission. Concentrations in the northern hemisphere are about 15% higher than in the southern hemisphere, and its atmospheric history has been monitored reasonably well since the 1960's. Surface water concentrations equilibrate with the atmosphere on a time scale of about 1 month. Present measurement techniques are substantially less tedious and time-consuming than for ^{39}Ar, but 250 to 500 liters of water are still required for each measurement. It appears that the low atmospheric krypton background will make the use of laser spectroscopy for ^{85}Kr measurements considerably more straightforward than for ^{39}Ar. With this technique, the sample requirements could be reduced to about 1 liter of seawater.

Recommendations

1. The incorporation of a useful range of geochemical tracers into a workable field program is inconsistent with the water sampling and sample density requirements of most physical oceanographic studies. WOCE should encourage a merger with current large-scale geochemical tracer studies such as TTO. To avoid duplication of effort, this might be done with a simultaneous 2-ship operation in which one ship would be used for physical oceanographic observations and small-volume high density geochemical measurements, and the second ship would keep pace with large volume but lower density specialized geochemical sampling.

2. WOCE should support the development and implementation of new analytical techniques which promise either to reduce logistic constraints on the density of geochemical observations or to improve the quality of the measurements.

3. Even for the most routine hydrographic and chemical measurements, it is essential that a highly qualified pool of technical support personnel and sampling and analytical facilities be established and maintained. This should include the upgrading of outmoded techniques for routine measurements and the improvement of data reduction and handling procedures. In the past, there has been frequent but unproductive discussion of establishing such a national capability. WOCE should provide the necessary incentive.

WOCE PLANNING DOCUMENT

Seasonal Variation and Water Mass Conversion

J.D.Woods

Prepared for the first meeting of the

WOCE SCIENTIFIC STEERING GROUP

Woods Hole, August 1983

First draft: 30 July 1983

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5. CONCLUSION

1. INTRODUCTION

The World Ocean Circulation Experiment (WOCE) is concerned with the collection of a data set that will serve to test and stimulate the development of coupled ocean-atmosphere models designed to predict climate change on the time scale of decades and longer (WCRP Stream 3). The WOCE data set will have two principal objectives:

1. Global circulation
2. Water mass conversion.

This paper has been written to open discussion on two questions that have arisen in preliminary consideration of the specification for WOCE. They are:

1. What aspects of the seasonal cycle in the ocean should be included in the WOCE data set ?
2. What large-scale signatures of water mass conversion should be included in the WOCE data set ?

It is convenient to combine the background material needed to discuss these two questions in a single paper because water mass conversion occurs principally in the boundary layer, where the seasonal cycle is the controlling factor. However, in considering the first question we shall also touch on aspects of seasonal variation related to the first objective of WOCE, circulation. And in considering the second question it is appropriate to include diabatic processes that effect water mass conversion below the boundary layer, since a knowledge of their regional variation may be essential for predicting how scalars are distributed around the World Ocean, and conversely for determining the mean circulation from observed distributions of scalars.

Seasonal and interannual variation

It is worth defining the expression "seasonal variation". It is normally taken to include the annual cycle and such of its harmonics as are needed to describe the mean seasonal cycle. The concept of a mean cycle poses difficulties. It can be defined experimentally in the context of a data set that samples the broad band of variability in the ocean. The phase is related to the astronomical cycle of insolation. It is convenient to discuss theories of the ocean boundary layer in terms of a hypothetical "mean" seasonal cycle, the inter-annual deviations from

it, and secular trends in the parameters defining the "mean" cycle. Such treatment can be rationalized by appealing to the notions of ensemble statistics and ergodicity. There are dangers in adopting such ideas to study the climate system with its broad spectral band of variability. Furthermore, there are reasons for believing that there can be no such thing as a regular seasonal cycle in the ocean. According to that conclusion, inter-annual variability in the boundary layer is not an unfortunate "climate noise", but an inevitable and essential feature of the seasonal cycle with important implications for water mass formation. In this paper, seasonal variation will be taken to include both the mean seasonal cycle and interannual variation produced by the same physical processes.

Water mass conversion

Water mass conversion also needs careful definition. Strictly speaking the temperature, salinity or individual concentrations of dissolved chemicals can only be changed by diabatic processes, including:

1. absorption or emission of radiation
2. molecular diffusion and conduction
3. chemical reactions (e.g. in the carbon cycle)
4. biological processes (e.g. oxygen, nutrients).

In practice, we do not deal directly with these diabatic processes, but with larger scale processes that are not themselves diabatic, but which control those that are:

1. Extinction coefficient
   (related to biological productivity)
2. Three-dimensional turbulent mixing
   (related to the various energy sources)
3. Mean change of chemical concentrations in large boxes
   (related to the distribution of turbulence)
4. Primary productivity
   (depending on (1), (2) & (3)).

The rate of water mass conversion is parameterized in terms of large-scale variables by means of models of the various processes involved. For WOCE we must focus on those larger variables and encourage the development of models linking them to the diabatic processes. Despite considerable progress in recent years, few of the theories of water mass conversion rest securely on experi-

mental evidence that can be applied with confidence at all depths and in all regions.

Regional variation

The aim of WOCE is to provide a global perspective in studies of ocean circulation and water mass conversion. However, the design of the experiment must take account of the large regional variations, between the hemispheres and ocean basins, and with depth. On the other hand it is important to minimize subdivision of the World Ocean. In discussing seasonal variation and water mass conversion it is sufficient to divide the ocean in the following ways:

Hemispheres	Northern and Southern
Oceans	Pacific, Atlantic, Indian, Mediterranean, Arctic, Antarctic
Layers	the seasonal boundary layer the cold water sphere the extra-tropical warm water sphere the tropical warm water sphere the benthic boundary layer the continental shelf

Not all of these sub-divisions will be discussed explicitly in this paper, but the need to cover them will have to be borne in mind in the design of WOCE. The allocation of priority to each region will reflect its intrinsic importance to the WCRP objectives (warm- versus cold-water sphere; seasonal- versus benthic boundary layer; continental shelf versus open ocean), the broader oceanographic aims of WOCE (meridional overturning versus wind-driven circulation; upper- versus deep-ocean water mass conversion), our present state of ignorance (southern versus northern hemisphere; Pacific versus Atlantic), and logistic contraints. It is also important to take note of the WCRP oceanographic strategy, in which activities in streams one and two are automatically assimilated into stream three. Thus the collection of a global description of the seasonal variation of surface temperature, sea ice, etc (Stream one) and the tropical ocean activities of the TOGA project (Stream two) do not need discussion in this paper beyond noting that they are being planned by other groups within the WCRP.

2. BACKGROUND

In this section we briefly survey contemporary ideas and observations concerning seasonal variation and water mass conversion, concentrating on general processes rather than regional characteristics.

Surface fluxes

The astronomical cycle of daily solar heating is the most important single factor determining the seasonal thermohaline cycle in the ocean. But the annual variation of the weather is also important. Clouds modulate solar heating; wind modulates surface cooling, evaporation, mixed layer turbulence and Ekman transport; precipitation offsets water loss by evaporation; air temperature and humidity affect longwave radiation, conduction and evaporation. They all exhibit seasonal variation that influences the seasonal cycle of mixed layer depth, temperature and salinity (fig.1).

Our knowledge of the seasonal variation of surface fluxes comes largely from the analysis of merchant ship meteorological observations (summarized by Budyko 1956; Bunker 1976; Hastenrath & Lamb 1977-79). The uncertainty in these estimates (probably 30-50 W/m^2) is an important source of error in diagnostic studies of water mass formation. It limits attempts to test models of the mean global circulations of heat and water (e.g. mean surface energy flux for the North Atlantic is ca. 50 W/m^2), and the thermal response of the climate system to CO_2 pollution (for which the predicted secular change of energy flux is ca. 1 W/m^2 per decade). Although the seasonal variation <u>per se</u> is not the main concern of WOCE, it is important to remember that substitution of the annual mean surface fluxes in models gives the wrong mean sea surface temperature and meridional circulation (Bryan 1982); just as neglecting the diurnal cycle of solar heating gives significant error in simulation of the seasonal cycle of surface temperature.

The seasonal cycle of wind stress is of central importance in the tropics, where the Ekman transport is large. Analysis of ship observations (O'Brien et al 1982) reveals the inter-annual

variation. The TOGA group is concerned with seasonal variation of surface fluxes in the tropics.

The seasonal boundary layer

The seasonal boundary layer is defined as the layer in which diabatic processes related to local surface fluxes control changes in temperature, salinity and chemical constituents of seawater occur sometime during the year. The depth of the seasonal boundary layer is determined by the annual maximum depths of the following seasonally-varying phenomena:

Depth	Phenomena	Comment
S	solar heating	threshold value 10 $MJ/m^2\,y$
D	buoyant convection	density gradient unstable
H	Turbulent buoyancy flux	density gradient stable

The limit is determined by convection at high latitudes, by solar heating at low latitudes, and by turbulent mixing along the equator (fig. 2). The depth of the mixed layer can be defined by the values of D and H at night, when they coincide. The seasonal boundary layer thus includes the annual cycle of heat and water storage resulting from fluxes through the surface. The diurnal and seasonal undulations of S, D & H leave the lower levels of the boundary layer (the seasonal thermostad and seasonal thermocline) free from surface-induced diabatic processes for much of the year. In regions of Ekman suction or where D increases downstream, the permanent thermocline water intrudes into the lower levels of the seasonal boundary layer each year, and is subsequently entrained into the mixed layer during the winter cooling season. The seasonal cycles of depth limits of processes in the boundary layer are shown schematically in fig. 3.

Seasonal variations in heat and water content

Gill & Niiler (1973) have presented theoretical arguments showing that, averaged over a large area (say 10 Mm^2), the annual cycles of heat and water content of the extra-tropical (latitude higher than 15°) ocean are dominated by the fluxes through the surface. The currents (geostrophic and Ekman) are not fast enough to contribute significantly on such large scales. The rather large uncertainties in surface fluxes calculated from merchant ship observations make it impossible to test that conclusion. However, a number of attempts have been made to estimate the climatologically important advective contribution as a residual (e.g. Stommel 1980). The WCRP Cage group (1982) assessed the feasibility of measuring the net annual surface heat flux to ± 10 W/m^2 over the North Atlantic ocean. Their proposals included two ingredients relevant to WOCE:

(1) direct measurement of the horizontal heat flux in the ocean by hydrographic coast-coast sections à la Bryden & Hall (1980),

(2) measurement of the change of heat content of the North Atlantic over the five year period of the Cage project by means of a ship-of-opportunity XBT programme. Bretherton, McPhaden & Kraus (1983) believe that the mean change in heat content can be measured in this way to ± 300 MJ/m^2y, despite the large annual cycle and mesoscale noise.

The JSC & CCCO have decided that these aspects of the Cage project should be incorporated into the WOCE.

Uncertainty in the distributions of precipitation and evaporation over the ocean makes it difficult to calculate the annual variation of water content. However, the seasonal variation of salinity profile, which reflects changes due to the net surface water flux, advection and mixing, can be used as an indicator. Fig.4 shows the variation of upper ocean water content along a 2.5 Mm batfish section between the Azores and Greenland, which is in remarkably good agreement with atlas data (Woods & Bauer 1983).

The steric change in sea level that accompanies the seasonal change in heat and water content of the upper ocean amounts to about 10 cm in middle latitudes, and can approach 50 cm in the tropics, where Ekman convergence dominates surface fluxes. The Japanese time series of hydrographic sections along 137°E provide a unique insight into the contributions of the seasonal (mean & interannual) variability of heat and water content to changes in sea level (Masuzara & Nagasaka 1983).

Seasonal variation in the mixed layer

Mixed layer depth

The classic description of the seasonal cycle of extra-tropical mixed layer depth is given by Tully & Giovando (1963), who analysed data collected at OWS "P". They introduced the idea of a potential layer depth H to take account of short term fluctuations (weather, diurnal). The seasonal variation of H represents the lower envelope of H', the instantaneous mixed layer depth. The seasonal cycle of the variance of (H-H') has a maximum in spring. We shall discuss its impact on water mass formation later. Meanwhile we concentrate on the seasonal variation of potential layer depth H.

Kraus & Turner (1967) showed that the astronomical cycle of solar heating is the major factor controlling the variation of H during spring, summer and autumn (fig. 5). Woods & Barkmann (1983) investigated the smaller contributions of the seasonal variation of weather . The winter variation of mixed layer depth depends on the penetration of density stratification established by advection as well as local heat storage, so the classic one-dimensional model (driven only by local surface fluxes) cannot be used to predict D the annual maximum depth of the mixed layer. The following argument suggests that there cannot be a regular seasonal cycle of mixed layer depth (i.e. constant D), even with a regular seasonal cycle of weather. A regular seasonal cycle would have to balance three budgets: heat, water and buoyancy. Surface fluxes of heat and water are not correlated; the seawater advected seasonally into the boundary

layer has a T-S relation that depends on that found below the boundary layer (e.g. central water) and thermoclinicity/baroclinicity in the seasonal thermostad (see below). The buoyancy constraint on mixed layer deepening hinders a solution for D that automatically closes the heat or water budgets in a single year. The winter depth (D), and the heat and water budgets of the seasonal boundary layer must exhibit inter-annual variability. (For further discussion see Woods 1983b.)

The measurement of D presents problems. It is not often that a research vessel happens to be at sea at the end of winter making a time series of precision CTD stations from which that with the greatest mixed layer depth can be unambiguously identified. Notable examples include the winter expeditions to the Mediterranean Sea (Medoc group 1970) and the Labrador Sea (Lazier 1973). Killworth (1983) has commented on the probability of encountering one of the "chimneys" where bottom water is formed. The criterion for determining mixed layer depth in such studies is that the profile should exhibit an adiabatic lapse rate to the accuracy of the instrument. Very few examples exist.

The more popular alternative is to examine profiles collected around the end of the cooling season for an elbow separating the well-mixed (past tense) upper layer from the thermocline below. Even if the station was occupied some days or even weeks after the moment of deepest mixing (i.e. when H=D), the elbow marking the bottom of the seasonal thermostad remains recognizable (fig. 6). Reid (1982) has advocated the use of the elbow in the oxygen profile in the same way (fig. 7).

In preparing atlases of climatological mean mixed layer depth (e.g. Bathen 1972; Robinson et al 1979; Levitus 1982) the usual criterion has been that H is the depth at which the temperature or density deviates from the surface value by some prescribed amount. Errors arise in estimating D by that method when a near surface thermocline formed in early spring calm sunny weather exceeds the prescribed temperature difference. Maps of D published by different authors show quite large differences (fig. 8).

A different approach has been introduced by Woods (1983b). D is identified with the haloclinicity elbow in a hydrographic section, rather than with the temperature/density/oxygen elbow in a single profile (fig. 9). The method works best with sections made in summer, which greatly increases the scope for systematic surveying. It depends on the fact that the seasonal change in density is due more to change of temperature than of salinity. Naturally, correction must be made for advection in strong currents, but that source of error is also present in the analysis of single profiles.

Mixed layer temperature

The seasonal variation of mixed layer temperature provides the principal ocean variation in the boundary conditions of atmospheric climate models. It is therefore the main oceanographic target of WCRP Stream one research. Plans are being made for monitoring the mixed layer temeprature globally. The existing data base is sufficiently rich in most the Northern Hemisphere to permit quite detailed descriptions of the regional variation of the mean seasonal cycle, and to make a start on describing inter--annual variations. The former are available in standard atlases (e.g. Robinson et al 1979; Merle 1978). EOF analyses of the inter-annual anomalies have been published by Barnett & Davis (1975), Weare (1977) and Navato et al (1981). Rodewald (1983) has drawn attention to the changes in annual mean temperature observed at the OWS (fig.10). Decadal variation of global mean mixed layer temeprature over the period 1861-1980 have been published by Folland & Kates (1983), who draw attention to the maximum at the end of the 1960s.

The Kraus-Turner model was designed to predict the seasonal cycle of mixed layer temperature from surface energy fluxes. The technique has been widley used in climatological studies of air-sea interaction (reviewed by Woods 1983a,b). Thompson (1976) showed that the inability of such models to predict winter mixed layer depth doe not have too serious an effect on the prediction of the annual cycle of mixed layer temperature. Kraus & Turner (1967) used the astronomical cycle of solar heating to force the model; Gill & Turner (1976) used OWS data; Woods & Barkmann(1983)

used Bunker's monthly mean climatological data (fig. 1). The deviations from observed mixed layer temperature are roughly what one would expect, given the uncertainty in the surface fluxes and neglect of advection ($^3/_4$ km.K/y at OWS "C"), and the net water flux (± 1 m). The seasonal cycle of temperature near the equator presents special problems that are being studied by TOGA.

Attempts to model inter-annual variations have been based on (1) slow relaxation to fast random forcing by the weather (Frankignoul & Hasselmann 1977), (2) forcing by equatorial waves (Hughes 1980), and (3) secular changes in surface IR due to CO_2 pollution (Bretherton 1982).

Mixed layer salinity

The seasonal variation of salinity is poorly documented. It is less easy to measure accurately than temperature and care has to be taken in selecting reliable data from the archives. Taylor & Stephens (1980) have analysed the mean seasonal cycle and inter-annual variabilitv at nine North Atlantic OWS (fiq.11). It is measured regularly along the Japanese section at 137°E, and a few other standard sections (see Tokyo 1981 conference report, WCP No.) Seasonal variation of salinity is included in the Merle (1978) and Levitus (1982) atlases, but the annual mean only is included in those of Robinson et al (1976, 1979). It is surprising that, despite the widespread use of recording thermo-salinographs on research ships, no attempt has been made to investigate the seasonal variability of salinity, as has been done so effectively with temperature. Of course, the dominant astronomical forcing which gives such a simple relationship between upper ocean heat content and mixed layer temperature (see fig. 12 , from Gill & Turner 1976), does not apply for salinity. The forcing is by precipitation and evaporation, plus run-off and sea ice freezing/melting in coastal and polar seas respectively, with significant modification by thermo-haline intrusions in the the seasonal thermocline at fronts. The equivalent water-content: mixed--layer-salinity curve will be more complicated, but worth exploring as a preliminary to designing a campaign for global monitoring of mixed layer salinity for the WCRP (as proposed recently by Bretherton.)

Modelling salinity changes in the boundary layer has tended to be relegated to the fringes of the World Ocean: to regions where there is sea ice and run-off, on the continental shelves and near estuaries, in peripheral basins (Arctic, Mediterranean & Red seas) and along the equator. However, the need to couple the global circulations of water in the ocean and atmosphere for WCRP stream three climate prediction brings open ocean salinity changes to centre stage. Levitus (1982) has drawn attention to the importance of salinity in determining the depth of winter convection. An accurate knowledge of inter-annual and decadal variation in the salinity source term for water mass formation is a central problem for WOCE.

Sea ice

The variation of sea ice is one of the greatest seasonal signals in the ocean climate system. The peak-peak variation of ice area is ca. 15 Mm^2 in the Antarctic (fig. 13). It has been mapped by satellite microwave radiometer for over a decade, allowing statistical analysis of inter-annual anomalies. Lemke (1982) has shown that they exhibit characteristics of a system that decays slowly after creation by random forcing (by the weather); the decay time of anomalies (a few months) is much shorter than the age of the oldest sea ice (several years). Thus the representation of sea ice in atmospheric climate prediction models is best handled by observed initial conditions, the anomalies in which decay to the mean cycle at the rate determined from multi-year observations. The effect of sea ice on air sea interaction comprises reductions in evaporation (by an amount that depends on the fractional coverage of leads and polynias) and in conduction through the ice (depending on its thickness, approaching zero as the ice reaches terminal thickness of two to three metres), and a sharp rise in albedo (when the ice is covered by snow). The presence of sea ice dramatically effects the winter climate of adjacent continents by hindering the formation of maritime air masses; it also fosters teleconnexions. Correlation between ice cover and variables associated with the global circulation of the atmosphere have been reviewed by Goody (1980). For these reasons, the monitoring of sea ice is one of the themes to be undertaken within WCRP Stream One, and may not need major investment by WOCE.

Models of sea ice (e.g. Hibler 1983) describe its seasonal growth and decay, and the changes of salinity in the mixed layer below it. The principal experimental focii are the complex processes occuring at the marginal zone (the MIZEX programme in the Arctic) and on the interpretation of microwave images of sea ice (microwave emissivity is changed by the freezing-melting cycle of multi-year ice). A possible sea ice project in the Antarctic is being discussed by WMO.

The seasonal variation of sea ice plays a significant role in the general circulation of the ocean. The boundary layer under the ice is controlled convectively by the salt and heat released at the freezing under surface, and by melt water run-off from the upper surface. The convective-advective balance in the boundary layer produces a halocline in the enclosed Arctic Ocean, which has not yet been modelled accurately by water mass models of the World Ocean (e.g. Bryan & Lewis 1979). High salinity is important for creation of dense water that ventilates the lower levels of the cold water sphere. In the Arctic it is supplied by inflow from the exceptionally salty North Atlantic; in the Antarctic it comes from the much higher rate of ice formation permitted by Ekman ventilation of ice away from the continent (Gill 1973). The association with bottom water formation with sea ice is consistent with evidence that the cold water sphere began to form before the polar glaciers achieved significant volume.

The seasonal thermostad

The vernal ascent of the mixed layer from its annual maximum depth D to the summer minimum occurs rapidly leaving a layer of weakly stratified water, the seasonal thermostad (fig.5). Deep convection during winter creates the thermostad; a combination of processes are responsible for its restratification during the rest of the year. First is the creation of a fossil thermocline during the ascent of the mixed layer in April and May; second comes solar heating below the mixed layer; third is geostrophic advection; and fourth is thermohaline instrusion. The relative importance of the four contributions in different regions can be assessed by inspecting climatological atlases in the light of one-dimensional models of the seasonal variation in the mixed layer. The advective contribution is known from the annual imbalance in surface fluxes. Being baroclinic it can also introduce an unstable salinity gradient which yields a buoyancy flux at the base of the mixed layer during the cooling season; a form of thermo-haline preconditioning of the thermostad for deep convection in winter (Killworth 1983; Woods 1983b). The depth of the seasonal thermostad is one of the key factors in calculating the annual rate of water mass conversion in the boundary layer. The subtle way in which it is restratified during the year determines that depth. Our knowledge of it is at present inadequate. We shall have to get to grips with it in WOCE.

Water mass conversion

The parameterization problem

As stated in the introduction, we are not concerned with the truly diabatic processes that occur at molecular scale, but with the microscale fluid motions that accelerate them, and hopefully with the mesoscale motions that in turn control the mean rates of water mass conversion effected by those microscale motions. The width of the spectrum from the Rossby radius (order 10 km) which might be resolved in observations (e.g. acoustical tomographic monitoring) and in water mass models of the World Ocean, to the molecular dissipation scale (order 1 mm) is dauntingly broad (seven decades) so it is not surprising that parameterization schemes designed to bridge the gap are still primitive. Only a few oceanographers are involved in searching for ways across the spectral knowledge-desert of the mesoscale and microscale. Instruments have been made that can measure microstructure down to the molecular dissipation limit, and others that can survey mesoscale finestructure, but experimental design is beset by problems of inadequate sampling that provide a lively subject for dispute in ocean turbulence circles.

There is no question of including such difficult observations of micro- & finestructure in the World Ocean Circulation Experiment. What is needed is a set of large-scale parameters that can be measured world wide during WOCE, and which will afterwards permit estimates of the regional variation of the rates of water mass conversion. The identification of such parameters implies that oceanographers are confident they know how to parameterize the processes in that seven decade spectral terra incognita. Despite notable advances during the past two decades, nobody in the business would make such a claim. It is more honest to admit that even the most confidently advocated ideas remain tentative, and that the best that can be offerred is a statement of contemporary ideas, with a suggestion that the word "parameters" be dropped in favour of the softer expression "signatures" of water mass conversion.

Water mass conversion in the seasonal boundary layer

For many purposes it is appropriate to treat all processes occuring below the boundary layer as adiabatic. Water mass conversion then occurs exclusively in the boundary layer. Attention focusses on the diabatic (or surrogate diabatic) processes in the boundary layer:

1. solar heating
2. buoyant convection
3. turbulent diffusion

and on the rate of subduction of seawater from the boundary layer into the quasi-adiabatic interior of the ocean. The properties of the subducted seawater are those of the mixed layer at the end of the cooling season when its depth is greatest that year. The volume of water subducted each year depends on three properties that exhibit important regional variation:

1. the horizontal circulation in the boundary layer
2. the annual maximum depth of convection
3. the annual vertical displacement by Ekman pumping

A detailed understanding of the seasonal variation in the boundary layer is a pre-requisite for estimating the properties and mass of water subducted at each region each year. The earlier conclusion that inter-annual variability is an intrinsic characteristic of the boundary layer depth and heat/water content, suggests that we should expect inter-annual variability in water mass formation beyond that attributable to inter-annual variation in surface fluxes:

Brief remarks on these processes and properties follow. For a more detailed review of the literature the reader is referred to the chapter on "The upper ocean and air/sea interaction in global climate" in Houghton (1983).

Solar heating

Solar heating is the most important diabatic process in the climatology of the World Ocean. It is not sufficient to know the energy flux at the surface: the vertical distribution of solar heating inside the ocean must also be known because it affects the depth of the mixed layer and heats water below D in the tropics. The first systematic investigation of the climatology of the vertical distribution of solar heating has recently been completed by Woods, Barkmann & Horch (1983). They have compared parameterizations of the vertical profile with observations, and studied the sensitivity of the profile to cloud cover and sea-water turbidity. They conclude that significant heating occurs below the mixed layer in the tropics, the rate being more sensitive to plankton concentration than cloud cover (fig. 2). This heat escapes to the atmosphere by upwelling into the tropical mixed layer or by advection to higher latitudes where the mixed layer gets deeper in winter.

The requirement is to measure the global distribution of solar heating to an accuracy commensurate with the change of net surface IR flux due to doubled CO_2, which Ramanathan (1981) gives as 100 MJ/m^2y. It is now possible to monitor the surface flux of solar energy to about 500 MJ/m^2y by satellite (Gautier 1982; Rashke 1983). It is also possible to measure seawater turbidity from satellite to within about one Jerlov water type, giving an uncertainty of $\pm$ 100 MJ/m^2y below the tropical mixed layer. Where the solar heating profile is known accurately it is possible to estimate the advective heating of the extra-tropical seasonal thermostad from bathythermograph data.

The appropriate "signature" is ocean colour index (Højerslev 1980) measured globally by satellite ocean colour scanner.

Buoyant convection

The characteristic of buoyant convection is thorough mixing of scalar anomalies, giving a negligible vertical concentration gradient in the familiar "mixed layer". Calculation of the depth to which convection penetrates in the ocean at any instant requires a knowledge of the surface buoyancy flux and the vertical density profile, and an assumption about the fraction of potential energy being released that is used in turbulent entrainment of the underlying pycnocline. Opinion is divided about the last point; some authors (e.g. Killworth 1980) suggest it decreases with depth of convection and may be negligible when deep in winter. If so the convection depth can then be calculated solely by adjustment. Solar heating during the day greatly reduces the depth of convection and the convective source of turbulent kinetic energy in the mixed layer (Woods 1980); neglect of this diurnal effect leads to error in calculations of the seasonal variation of mixed layer depth and temperature.

Turbulent diffusion

The sharp decrease in depth of the convection layer during the morning (due to solar heating, see above) leaves a weakly stratified diurnal thermostad (fig.3). There turbulence generated by the wind stress and (inertial & geostrophic) current shear diffuses scalars down gradient and changing their concentration C according to the curvature $\dot{C} = d_z(Kd_zC)$. The vertical distribution of K can be calculated from a one dimensional model based on parameterization established experimentally in laboratory experiments (Mellor & Yamada 1982). The surface wind stress and buoyancy flux must be known.

Normally it is not necessary to simulate the diurnal variation. It is sufficient to follow the seasonal change. In which case convection will homogenize the diurnal thermostad during the following night, and the task of the model is to calculate the maximum depth of the mixed layer each day. The most important exception is the equatorial undercurrent, where strong turbulence persists day and night below the convection layer, giving a climatologically significant vertical heat flux (fig.14).

Water mass subduction

We distinguish between two kinds of subduction: horizontal and vertical.

The critical factor in the former is a negative downstream gradient of D (the annual maximum depth of the mixed layer), so that water mixed to great depth by convection one location flows geostrophically under the boundary layer downstream during the following winters. The extreme example of that process is the formation of bottom water at chimneys of convection extending to the ocean floor over a small area (order 100 km across). But it is important more generally in the formation of deep water. The signature is $\underline{U}.\nabla D < 0$.

Vertical subduction is driven by Ekman pumping in the classic manner envisaged by Iselin (1939). Stommel (1979) has described the process in the presence of a seasonally varying mixed layer depth, showing that (as with horizontal subduction) it is the winter properties of the mixed layer that are carried down. The seasonal variation of the mixed layer depth (H) and the annual range of vertcial displacement by Ekman pumping must be known. In WOCE the latter might be determined by global monitoring of wind stress by satellite scatterometer.

Seasonal variation of seawater properties below the boundary layer

Wüst (1935) pondered the possibility of encountering in the deep ocean a banded structure in seawater properties recording their seasonal variation in the mixed layer from which they travelled down, as though along a continuous inclined conveyor belt. We now know that the process of water mass formation -the addition of new water from the boundary onto the conveyor belt (which we now tend to associate with an isopycnic surface) is quantized annually by the seasonal variation of mixed layer depth. Nevertheless it is worth examining the downstream variation of scalar properties on upper ocean isopycnals in order to learn about the sequence of winter events in which seawater that acquired its properties by diapycnic processes in the boundary layer became subducted into the quasi-adiabatic interior. For a recent example for Jenkins (1982).

Water mass conversion below the boundary layer

Until quite recently the ocean was believed to be continuously turbulent at all depths with the intensity, and therefore the eddy diffusivity modulated according to the Richardson number. But flow visualization studies in the 1960s established that below the mixed layer, the flow is predominantly (95%) laminar. Turbulence was observed to occur in the seasonal thermocline as short-lived isolated events resulting from shear instability in internal waves. That discovery led to the present paradigm in which the natural state of the flow is laminar at all depths below the mixed layer. Molecular diffusion occurs universally and is enhanced locally by rare, transient microscale events, the accumulated effect of which is nevertheless greater than that of the universal molecular diffusion.

It is these events therefore that are held to be responsible for water mass conversion below the boundary layer. They are weak and, as we noted above, for many purposes it is not inappropriate to ignore them entirely and to assume that the processes in the interior of the ocean are adiabatic. Nevertheless, diagnostic studies of observed distributions of tracers leads to the conclusion that there is diffusion in the deep ocean at a rate faster than molecular.

Because mixing by quasi-geostrophic eddies rapidly disperses scalars along isopycnic surfaces it is only necessary to discuss the mixing of scalars by these microscale events in the cross isopycnic direction. If the mixing is parameterized by a Boussinesq eddy diffusivity, then we distinguish between an isopycnic diffusivity, whose value depends on the kinetic energy of mesoscale eddies, and a diapycnic diffusivity, whose value depends on the net effect of intermittent microscale events. Here we shall concentrate on just two types of microscale event: billow turbulence, and double diffusive convection. The effects of cabbeling and of intrusion from the benthic boundary layer will not be discussed.

Billow turbulence

The first process involves the mechanical overturning of isopycnals by billows resulting from internal wave induced shear instability. The mechanism was discovered in the seasonal thermocline (Woods 1968) and it is believed to occur at all depths. The loss of energy from the billow turbulence as it does work against the Archimedes force is well understood from laboratory experiments, but translation of that into a widely valid parameterization for diapycnic eddy diffusivity requires a statistical theory for the frequency of occurrence and size of the billows in terms of their energy source, the internal waves. So far no satisfactory theory has been developed (Munk 1981). The problem arises from uncertainty concerning the physical processes involved in cascading energy across the spectral gap between the shortest wave (order 10 metres) and the largest billow (order 10 cm). The evidence from the seasonal thermocline is that density fine structure plays a role. Some of the fine structure is produced by internal wave strain, but some is produced by mesoscale turbulence and thermohaline intrusions that are not controlled by internal waves. So it looks as if a completely closed theory, based solely on internal waves, may be physically unrealistic. But it is not clear how relevant the evidence of the seasonal thermocline is to the deeper layers of the ocean. Attempts to resolve that question by examining microstructure profiles have proved inconclusive. Meanwhile, the working hypothesis of billow turbulence theoreticians is that for most of the World Ocean, however complex the physical processes at the spectral gap, they can be parameterized to sufficient accuracy in terms of the Brunt-Väisälä frequency. Garrett (1983) has reviewed the results of recent work carried out in that spirit.

Microstructure measurements in the main thermocline have been interpreted in terms of billow turbulence, but there is disagreement about the statistical interpretation of the data, which depend on rather small samples of a highly intermittent process. Further experiments are needed before such data can provide the evidence needed to test parameterizations based on speculative theories. There is a reasonable chance that the situation will become clearer during the five years before WOCE. Meanwhile, planning should proceed on the basis that N will provide a sufficient signature of water mass conversion due to billow turbulence.

Double diffusive convection

The predominance of laminar flow in the ocean provides the opportunity for differences in molecular diffusivities for heat and salt to provoke convection in midwater, when one constituent is statically unstable. Extensive laboratory studies (reviewed by Turner 1973, 1981) have provided the framework for interpreting steplike structures in CTD profiles in terms of double diffusive convection. Recent work by Schmitt (1981) has established the fact that the process procedes at a rate governed by the density ratio ($R = \alpha T_z/\beta S_z$), and acts to adjust the T-S profile to a universal form with R approximately 2. It is as though double diffusive convection "diffuses" R, removing any anomalies that might have originated during water mass formation (fig.15). This one dimensional theory works well in the central water. Clearly the relevant signature for WOCE is the density ratio R.

Double diffusive convection plays an additional role at strongly thermoclinic fronts, such as the North Atlantic polar front. At sites along the front where mesoscale frontogenesis or curvature isopycnically creates a small fold in the otherwise doubly stable water, the conditions for double diffusive instability are met, and the fold is rapidly extended by the lateral pressure gradient created following release of the latent potential energy of the now unstable salinity gradient (fig.16). Such cross-front intrusions have been identified in many surveys of mesoscale finestructure. Parameterization in terms of frontal baroclinicity (for the initial fold) and thermoclinicity (for the subsequent intrusion) can be expected shortly. The process is a form of sloping convection, which draws mainly on the thermoclinic available potential energy at the front. Consideration should be given to measuring the thermoclinicity and baroclinicity at major fronts as part of the WOCE.

Seasonal variation in circulation

The survey by Gill & Niiler (1973) provides a starting point for discussion of the seasonal cycle in dynamical variables, and rather than attempt to summarize the conclusions of that paper I have included the authors' abstract as appendix 1 below. They were concerned with the largest scales of variability (5°x5°) at latitudes higher than 15° and away from boundaries.

Attention is drawn to their discussion of seasonal variation of sea level. They note the relatively large inverse barometer effect (a few centimetres) at high latitude; the small barotropic response (a few millimetres) to seasonal changes in wind stress; and steric changes (a few centimetres) due to seasonal change in heat content due to (a) net surface fluxes, (b) Ekman fluxes in the mixed layer, and (c) vertical displacement of the ocean thermocline by seasonal variation of Ekman pumping. The Sa tidal component (a few millimetres) is discussed by Cartwright (this volume, chapter).

Gill & Niiler (1973) raise the question of advective contribution to the extra-tropical seasonal variations of heat and water content, which is negligible according to their dynamical analysis, but assumed to close the significant gaps in the surface annual energy and water budgets. They rejected Bathen's (1971) interpretation of the seasonal cycle heat content in terms of advection and (horizontal) eddy diffusion. They note the importance of Ekman advection in the seasonal variation of heat content at low latitudes. Bryan (1982) has shown that seasonal variation of Ekman transport can produce the inter-hemispheric exchange deduced by Oort & Vonder Haar (1976) as a residual in their analysis of atmospheric heat storage and transport and oceanic storage. The TOGA group (1983) summarize the important seasonal variations occuring in the tropical ocean, where the large Ekman transport and fast baroclinic waves influence heat content as much as surface fluxes in many regions, and where inter-annual variation often exceeds the mean seasonal range.

Although the wind-driven gyres are not expected to exhibit significant seasonal variation, there have been reports of seasonal modulation of mesoscale variability. The NEADS group (1981) have produced evidence of seasonal variation in fluctuations with periods of a few weeks. Baroclinic planetary waves with annual period are believed to radiate across the Pacific from seasonal-upwelling regime along the West coast of America (Mysak 1983). Inter-annual modulation of the upwelling by waves propagating along the coasts from the equator may find a response in these open ocean Rossby waves. Seasonal variation in the Florida current (Niiler & Richardson 1973; Schott 1983) is characterized by inter-annual variation larger than the mean cycle.

3. THE WORLD OCEAN CIRCULATION EXPERIMENT

In this section we identify WOCE requirements for measurements of the seasonal cycle and water mass conversion, note the existence of other projects that may supply the required data, and propose actions to collect the rest, either within WOCE or as subsidiary projects.

WOCE requirements

Water mass conversion (wmc) is one of the central themes of WOCE. It is believed that after computers become sufficiently powerful to permit adequate resolution of topography and mesoscale variability in the ocean the key problem facing ocean climate modellers will be parameterization of processes that effect water mass conversion. The only reason that wmc is not given top billing now is that it is more urgent to reduce errors in the dynamics due to inadequate resolution that tend to mask those due to inadequate parameterization of wmc. But the situation is likely to change soon; at least before delivery of the WOCE data set.

The first decision to be made is whether it would be sufficient for WOCE to concentrate exclusively on the seasonal boundary layer, where water masses are formed, or whether it is also necessary to include in the WOCE data set some indicators of the regional variability of processes that change water mass characteristics below the boundary layer. The argument for concentrating on the boundary layer is that water mass formation represents a more urgent problem of direct relevance to surface fluxes, to the coupling between ocean and atmosphere, and to the WCRP. To take a specific example, the calculation of oceanic scavenging of CO_2 pollution and the thermal response of the ocean to the CO_2 left in the atmosphere depend on accurate calculation of the rate of subduction of seawater from the boundary into the quasi-adiabatic interior of the ocean below. It is unlikely that neglect of diapycnic mixing below the boundary layer will seriously alter model predictions of climate change due to CO_2 pollution if they deal accurately with water mass formatiuon in the boundary layer of the upper ocean.

The case for including signatures of deep water mass conversion rests more on the other objective of WOCE, namely the general circulation. There are two aspects. The first concerns the diagnostic modelling of ocean circulation from distributions of scalars measured during WOCE (and, where appropriate, during earlier expeditions). There is evidence from diagnostic studies (McDougall 1983) that diapynic mixing has to be taken into account. Uncertainty in its nature and magnitude as functions of region may lead to error in the diagnosed circulation. An independent measurement of the diapycnic processes might help to constrain the model and yield a more precise estimate of the circulation - provided it is accurate.

The second case comes from the more general aims of WOCE, namely to provide a data set that will stimulate the development of and test general circulation models of the World Ocean. The equations of such models include terms to represent the net effects of diapycnic mixing events. There have been problems in

the past because of contamination by the diapycnic component of the intense lateral mixing due to quasi-geostrophic (and adiabatic /isopycnic) eddies. However, it is now known how to avoid that error by using isopycnic coordinates (Bleck & Boudra 1982) or by rotation of the isobaric coordinates (Redi 1982). Having overcome that problem, the question facing modellers is how to parameterize diapycnic mixing in terms of variables resolved in the model, and how to ensure that they vary realistically with depth and region. Field data are needed to help solve that question; they need to have the global coverage that is the hallmark of WOCE. But is it necessary to collect them during WOCE ? It is unlikely that any measurements we can contemplate making on a global scale will be precise enough to permit discussion of large-scale temporal variability below the boundary layer, so there is no case for simultaneity. The best argument for collecting such data during WOCE, rather than as a part of a separate programme, is the logistic one: research ships may not be going again to such out-of-the-way places for a long time. The argument against collecting data relevant to deep wmc during WOCE is that it may compete unacceptably with other WOCE objectives. In view of the present state of uncertainty about methods of parameterizing diabatic mixing events and the controversy over measurement techniques, the best policy for WOCE is to incorporate measurments related to wmc below the boundary layer on a non-interference basis, in the hope that they might yield a valuable bonus afterwards.

That relaxed approach is not recommended for water mass conversion in the seasonal boundary layer. Accurate parameterization of the processes involved are vital to the success of WOCE as a component of the WCRP and, more generally, for modelling the World Ocean water masses. Simultaneity is crucial because of the known inter-annual and inter-decadal variability in the upper ocean.

Seasonal variation in circulation

The reader is reminded that in this paper seasonal variation embraces not only the "mean" seasonal cycle derived from a multi-year data set, but also the attendant inter-annual variability. The requirement can be conveniently divided into two components:

1. global information of seasonal variability in ocean circulation and statistics of mesoscale variability
2. Studies in regions of special interest:
   e.g. (1) the tropics, where Ekman forcing is strong and baroclinic waves travel fast,
   (2) straits and sills crucial to the general circulation, including:
   the Drake passage
   the Florida straits
   the Straits of Gibraltar
   the Greenland-Iceland-Faroe-Scotland sill

Global monitoring will be effected by satellites and drifters. Satellite altimetry and Argoss drifters will provide the seasonal variation in surface geostrophic flow needed for modelling the boundary layer. It will be important to develop a statistical description of seasonal variation in the core of major currents (e.g. at the polar fronts separating subtropical anticyclonic and subpolar cyclonic gyres) the displacement of which may be the main cause of observed interannual (Eulerian) variations of surface temperature and salinity. Care will be needed to define the mean core of the current in the presence of intense meandering with shallow geostophic jet streaks. The seasonal variation of mesoscale eddying presents a challenge for WOCE in its own right. Having been detected at a few moorings, the challenge is now to use altimeter and drifter data to monitor the seasonal variation globally. Deep (acoustically-tracked drifters will help to establish the vertical distribution of seasonally-varying mesoscale eddy kinetic energy. Interpretation of the data will require a supporting programme to measure:

1. upper ocean heat content (steric change in sea level)
2. moored current meters (baroclinic structure of eddy KE)
3. ship of opportunity current profile programme

Global measurements of seasonal variation in Ekman transport and pumping will be derived from satellite scatterometry.

Seasonal variation of circulation (in the broad definition adopted here) is the quintessence of the TOGA programme. The equatorial region is less important when viewed from the global and decadal perspective of WOCE. However, the annual flow of heat from the summer to the winter hemisphere, identified by Oort & Vonder Haar (1976) and modelled by Bryan (1982) is an important feature of the global circulation and requires careful monitoring of the Ekman heat flux in the tropics. The monitoring of seasonal variation of mass transport and heat flow at straits and sills is also an important target for WOCE, and require intense local sub-programmes incorporating techniques that are not appropriate for global deployment, e.g. moored current meters, tomographic arrays, and repeated surveys by research ships.

Seasonal variation in the boundary layer

Calculation of water mass formation requires a detailed description of seasonal variation in the boundary layer, including its depth, temperature, salinity, vertical motion (Ekman pumping), advection (Ekman and geostrophic flow). These variables are needed globally to calculate the wide variety of water properties being subducted annually. (It is not clear that the traditional simplification in terms of classical water types named in Wüst 1935, and Sverdrup, Johnson & Fleming 1942 is appropriate for decadal climate prediction.) But it has proved troublesome to achieve an adequate empirical description of seasonal variation even at easy sites like OWS "P" (Tully & Giovando 1964). More general descriptions of regional variation include widely different estimates of such crucial properties as mixed layer depth in winter. Special efforts are needed to resolve these difficulties by careful in situ measurments at a small number of locations chosen as being specially relevant to WCRP rather than oceanographically tame. It is recommended that some such subsidiary programme be undertaken in support of WOCE.

The WOCE requirement for <u>global</u> coverage cannot be met economically by such in situ campaigns. Yet the global measurements planned for WOCE do not provide the information needed directly to describe seasonal variation in the boundary layer. Woods (1983a) proposed that that gap be bridged by using a model of the boundary layer. Satellite data (scatterometer, radiometers operating in the IR, visible & microwave) can be used to estimate monthly mean surface fluxes of solar energy, sensible and latent energy, precipitation, and Ekman tramsport & pumping. Geostrophic mean flow and mixing can be estimated from satellite altimetry and oceanic circulation models. Turbulent kinetic energy supply to the mixed layer can be determined from the wind stress and surface buoyancy flux. The problem with this approach at present is that the surface fluxes are likely to have quite large errors, which produce corresponding errors in boundary layer structure, according to sensitivity studies with boundary layer models. Furthermore, the latter do not yet achieve accurate simulation of the observed mean seasonal cycle from observed mean surface weather. For example, fig. shows a comparison of Robinson data and model prediction based on Bunker data. The present state-of-the-art is not too encouraging, but no other approach is on offer, so every effort should be made to collect the global observations needed, and to undertake pre-WOCE studies aimed at improving the performance of boundary layer models. It would also be worth investigating the potential of instrumenting a large number of Argoss-type buoys with thermistor/conductivity chains. Ship-of-opportunity campaigns should look beyond XBTs to surface thermo-salinographs, XCDTs and Doppler sonar current profilers in order to accumulate data that will test new boundary layer models.

<u>Water mass conversion in the boundary layer</u>

The seasonal variation in the boundary layer provides the key to calculating the annual rate of water mass formation region by region. Particular attention must be paid to the following variables.

1. The annual vertical displacement by Ekman pumping/suction,
2. The annual maximum depth D of the mixed layer.

The horizontal gradient and interannual variability of D are particularly important for calculating sub-polar water mass formation rates. D cannot be derived from existing boundary layer models (because of the, as yet unknown, interaction between advection and buoyant convection), but it can be measured in winter with high quality hydrographic stations, and in summer with oxgen profiles and hydrographic sections.

Sea ice

The seasonal variability of sea ice cover is important for the polar salinity balance and bottom water formation. It should be monitored throughout WOCE by satellite microwave radiometer. Independent programmes such as MIZEX designed to improve our understanding of the role of sea ice in water mass conversion should be encouraged as supporting the objectives of WOCE.

Water mass conversion below the boundary layer

Recent investigations relate diapycnic mixing due to wave-induced shear instability to the mean Brunt-Väisälä frequency N, and double diffusive adjustment of the T-S relationship to the density parameter R. Further research is needed to test and develop these ideas. It would be appropriate to invite SCOR WG69 ("Ocean turbulence") to help coordinate a pre-WOCE programme of microstructure case studies with that aim. It seems inappropriate to specify additional measurements for WOCE to estimate the rates of water mass conversion in deep water. If the above conjectures prove right, the necessary information will be found in the WOCE hydrographic section data.

Mixing at large scale fronts

The leakage of scalars across large-scale oceanic fronts is one of the more important processes governing their global distribution. Also, the vertical circulation occuring in the meso-scale jet streaks imbedded in large-scale fronts provides an important pathway between the mixed layer and quasi-adiabatic interior of the ocean for nutrients and other scalars. The polar fronts separating the subtropical anticyclonic gyres from the sub-polar cyclonic gyres are particularly important examples. The role of seasonal variation in scalar transport through these fronts is not known. Some attention should be given to monitoring them in the WOCE.

Other projects

In planning the WCRP it is necessary to take account of other projects planned within the WCRP or otherwise. The WCRP is divided into three parts, called streams and concerned respectively with climate prediction on time scales on weeks, years and decades. The oceanographic programmes for each stream are designed to be downstream compatible. Thus measurements to support stream one are also used in streams two and three; those in stream two are also needed for stream three. WOCE is part of stream three and benefits from oceanographic measurements in streams one and two.

WCRP stream one

The main oceanographic concern in stream one is with the measurement of the global patterns of surface fluxes of energy and water for use as boundary conditions in atmospheric climate models. An initial global monitoring programme for sea surface temperature, sea ice and possibly a few other surface variables will be needed to establish the mean cycle and statistics of interannual anomalies, including their average rates of decay.

WCRP stream two: TOGA

The main activity in stream two is the TOGA (Tropical Ocean and Global Atmosphere) project. TOGA will cover all aspects of seasonal variability and water mass conversion relevant to WOCE in the tropics. It is recommended that WOCE should not seek to duplicate measurements planned for TOGA. The two programmes should be integrated in the tropics, which cover nearly half the World Ocean. As a general policy, the WOCE scientific steering group should identify measurements required in the tropics and seek to have them included in the TOGA programme. It seems unlikely that that policy would lead to any unwelcome extras for TOGA with respect to seasonal variability and water mass conversion.

The WCRP Cage project

Originally planned as a separate programme to improve measurements of the surface energy flux on the scale of an ocean basin, the WCRP Cage project has now been subsumed within WOCE. Estimates of mean surface fluxes will be derived from transocean heat flux measurements, and from the external budget method pioneered by Oort & Vonder Haar. A ship-of-opportunity XBT programme is needed to determine changes in the upper ocean heat content in support of those aspects of the Cage programme. An attempt will be made to increase the accuracy of energy flux estimates from the bulk method by using systematic satellite data (scatterometer & IR radiometer). WCRP radiation experiments (ERBE & ISCCP) are significant contributors to these Cage activities. The question of whether all components of Cage need to be made simultaneously with WOCE is open for discussion.

4. PROPOSED ACTIONS

We now make a first attempt to see how the WOCE requirements with regard to seasonal variation and water mass conversion can be achieved, taking into account measuring systems tentatively identified as being available for WOCE and associated activities. It is convenient to introduce a classification of activities.

Class 1	Systematic global monitoring by satellite
Class 2	Random global monitoring by drifter
Class 3	Ship of opportunity campaigns
Class 4	Coast-coast sections by research ships
Class 5	Monitoring at critical sites (straits, sills, fronts) by moored instruments, tomography, etc
Class 6	Pre-WOCE studies (a) modelling (b) local field experiments
Class 7	WOCE monitoring of longer term variability
Class 8	Other WCRP experiments (e.g. TOGA, Cage, SECTIONS, PATHS, MIZEX, etc)
Class 9	Other non-WCRP activities (e.g. WWW, ECMWF)

The proposed actions are presented under five headings:

1. Seasonal variation of surface forcing
2. Seasonal variation of circulation
3. Seasonal variation in the boundary layer
4. Water mass conversion in the boundary layer
5. Water mass conversion below the boundary layer.

Seasonal variation of surface forcing

(see also WOCE position paper by Crease)

1. Wind stress

Class 1	Scatterometer
Class 6b	Calibration of scatterometer
Class 8	TOGA for tropics
	SECTIONS in energetically active zones
Class 9	WWW synoptic ship observations

2. Heat flux

Class 1	Scatterometer for wind in bulk formulae
	IR radiometer for surface temperature
Class 2	In situ measurements of surface temperature
Class 3	Dual IR radiometer for surface heat flux
Class 6a	Atmospheric boundary layer models
Class 6b	Development of the dual radiometer
Class 8	Cage
	TOGA in the tropics
	SECTIONS in energetically active zones
Class 9	Global weather analysis by ECMWF
	OWS and merchant ship reports

3. Evaporation

Class 1	Scatterometer for wind in bulk formulae
	IR radiometer for surface temperature
Class 2	In situ measurements of surface temperature
Class 6a	Atmospheric boundary layer models
Class 6b	HEXOS for parameterization at high winds
Class 8	TOGA in the tropics
	SECTIONS in energetically active zones
Class 9	HEXOS

4. <u>Precipitation</u>

Class 1	Microwave
Class 5	Island rain gauges
Class 6b	Calibration of satellite measurements
Class 8	TOGA in the tropics
	SECTIONS in energetically active zones
Class 9	Global weather analysis by ECMWF

5. <u>Solar energy</u>

Class 1	Visible radiometer
Class 6	Further development of parameterizations
Class 8	ISSCP

Seasonal variation in circulation

1. Gyre-scale circulation, basin modes, etc

Class 1	Altimeter
Class 2	Surface drifters tracked by satellite
	Deep drifters tracked acoustically
	Deep drifters that pop up for satellite fixes
Class 5	Monitoring the Florida current (in progress 1983)
	Monitoring the Drake passage
	Monitoring the Greenland-Iceland-Scotland gap
Class 6b	Florida current (see 5 above)

2. Fronts between gyres

Class 5	Polar fronts by tomography
Class 6b	North Atlantic polar front now being explored by research ships

3. Mesoscale variability

Class 1	Altimeter
Class 2	Surface drifters tracked by satellite
	Deep drifters tracked acoustically
	Deep drifters that pop up for satellite fixes
Class 6b	Analysis of long term mooring data (e.g. NEADS)

4. Seasonal planetary waves

Class 1	Altimeter
Class 3	XBT campaign (as used in analysis by Magaard, Mysak)
Class 5	Monitoring Eastern boundary source regions
Class 7	Annual cycle is only part of broader spectrum
Class 8	TOGA on the equator and tropics (El Nino)

Seasonal variation in the boundary layer

1. Heat content

Class 1	Altimeter (steric component)
Class 2	Surface drifters with thermistor chains
Class 3	XBT campaign
Class 5	Along the equator (see TOGA below)
	Special study in north Atlantic for Cage
Class 7	Decadal monitoring by tomography to detect secular trend due to CO_2 pollution
Class 8	TOGA in the tropics
	SECTIONS in energetically active zones

2. Freshwater content

Class 2	Surf. drifters with temp./conductivity chains
Class 3	XCDT campaign
Class 5	Along the equator (see TOGA below)
	Special study in north Atlantic for Cage
Class 6b	Batfish sections in north Atlantic
Class 7	Decadal variability is virtually unknown
Class 8	TOGA in the tropics
	SECTIONS in energetically active zones

3. Mixed layer temperature (N.B. WCRP stream one activity)

Class 1	IR radiometer; microwave radiometer
Class 2	Surface drifters
Class 3	Ship engine intake (N.B. needs calibration)
Class 5	In the tropics (see TOGA below)
Class 6	Establish relationship between different methods of measurement
Class 7	Part of WCRP stream one
Class 8	WCRP stream one activity globally
	WCRP stream two (TOGA) activity in tropics
Class 9	World Weather Watch ship observations

8. <u>Vertical motion</u> (from Ekman transport, 7 above)

9. <u>Sea ice</u> (N.B. WCRP stream one activity)

Class 1	Microwave radiometer
Class 2	Satellite tracked drifters on sea ice
Class 5	Arctic and Antarctic sites to be chosen
Class 7	Important to establish secular change e.g. due to CO_2 pollution
Class 8	WCRP sea ice programme now being discussed MIZEX

Water mass conversion in the boundary layer

1. <u>Solar heating</u>

Class 1	Visible radiometer (see "solar energy" above) Seawater turbidity from ocean colour monitor
Class 4	Solar irradiance profiles
Class 6a	Develop parameterization based on satellite data
Class 6b	Field work to support 6a
Class 8	TOGA in the tropics

2. <u>Annual range on Ekman pumping/suction</u>

(from "vertical motion" above)

3. <u>Annual maximum depth of mixed layer</u>

(See also "mixed layer depth" above)

Class 2	Surface drifters with thermistor chains
Class 3	XBT campaign
Class 4	Oxygen profile for Reid's method Thermoclinicity for Woods's method Winter profiles, if there are any
Class 5	Annual sections across sites of central water formation to use Reid/Woods methods
Class 6	Develop the above methods of determining the depth of winter mixing from summer data
Class 8	TOGA in the tropics

4. Mixed layer salinity

Class 1	Passive microwave (?)
Class 2	Surface drifters with conductivity
Class 3	Thermosalinograph; bottle sample calibration
Class 5	Tropics by TOGA (see below)
Class 7	At selected stations (OWS ?)
Class 8	TOGA in tropics

5. Mixed layer depth

Class 2	Surface drifters with thermistor chains
Class 3	XBT campaign
Class 5	Ocean Weather Stations
Class 6a	Develop boundary layer models that can-predict mixed layer depth
Class 6b	Support 6a with field experiments
Class 7	Investigate space-time statistics of the annual maximum depth of the mixed layer. (see below "water mass conversion in the BL")
Class 8	TOGA in the tropics

6. Geostrophic transport

Class 1	Altimeter
Class 2	Surface drifters tracked by satellite
Class 3	Doppler acoustic current profiler
Class 4	Doppler acoustic current profiler diagnostic modelling of WOCE data set
Class 5	Polar fronts (tomography ?)
Class 8	TOGA in the tropics

7. Ekman transport

Class 1	Scatterometer
Class 5	In the tropics: see TOGA below
Class 6a	Explore consequences of aliasing weather
Class 6b	Field work to support 6a
Class 8	TOGA in the tropics
Class 9	Global analysis of surface wind stress by ECMWF and other global meteorological centres

Water mass conversion below the boundary layer

1. **Billow turbulence due to internal waves**

Class 4	Profile of Brunt-Väisälä frequency if K=K(N)
Class 6	Invite SCOR WG69 to organise a programme of microstructure and finestructure case studies to establish parameterizations

2. **Billow turbulence due to geostrophic shear**

Class 5	Equatorial undercurrent
Class 8	TOGA

3. **Double diffusive convection**

Class 4	Density ratio (as used by Schmitt)
Class 5	Sections along subduction trajectories in the Warm water sphere (central water)
Class 6	Further experimental studies to establish parameterization of diabatic fluxes associated with double diffusive convection (SCOR WG69 ?)

4. **Double diffusive intrusions at fronts**

Class 5	Baroclinicity & thermoclinicity at major oceanic fronts (see also "fronts between gyres" above)
Class 6	Experimental studies to develop parameterizations of the transport effected by double diffusive intrusions in terms of frontal baroclinity and thermoclinicity (SCOR WG69)

5. **Benthic boundary layer** (No activity within WOCE)

6. **Continental shelf processes** (No activity within WOCE)

5. CONCLUSION

This paper has briefly reviewed the state-of-the-art in two themes relevant to the World Ocean Circulation Experiment, namely seasonal variation and water mass conversion. WOCE requirements with respect to both those themes have been identified, and proposals have been made about how to satisfy those requirements, taking account of existing ideas for a WOCE observing scheme, and other related experiments.

Seasonal variation has been taken to include both the mean cycle (comprising the annual wave and its harmonics) determined from a multi-year data set, plus the attendant inter-annual variability. Seasonal variation in circulation (including meso-scale mixing and planetary waves) and in thermohaline and chemical constituents of seawater both represent significant factors in planning the WOCE. So does seasonal variation in the surface fluxes, both because they force such changes, and also because they will be needed to test coupled ocean-atmosphere models of the planetary climate system. Many aspects of these seasonal variations can be monitored directly by observing systems proposed for WOCE, but pre-WOCE research is needed to improve methodology. The following cases present the most difficult problems and deserve special attention:

Surface fluxes
1. evaporation and sensible heat flux
2. precipitation
3. total heat flux (from dual IR radiometer on ships-of-opportunity)

Boundary layer
1. freshwater content from ships-of-opport.
2. surface salinity from satellite microwave
3. mixed layer depth, especially in winter

Water mass conversion is one of the central themes of WOCE. The strategy is to collect large-scale signatures of water mass conversion, appropriate for future use in diagnostic models of the general circulation and in prognostic models of climate change. Advances made during the past two decades makes it possible to identify prime candidates for such large-scale signatures. The following have been discussed in this paper:

In the seasonal boundary layer of the ocean

1. the vertical profile of solar heating
2. the annual vertical displacement by Ekman pumping
3. the annual maximum depth of the mixed layer

Below the boundary layer

1. the vertical profile of Brunt-Väisälä frequency
2. the density ratio (as in the T-S diagram)
3. frontal baroclinicity and thermoclinicity

Further research is needed to establish useful parameterizations based on these large-scale signatures. It is recommended that SCOR WG69 ("ocean turbulence") be invited to help by organizing a series of case studies involving microstructure and finestructure measurements at sites where the WOCE large-scale signatures are being monitored. Highest priority should be given to processes that effect water mass conversion in the seasonal boundary layer of the ocean, second priority to those in the warm water sphere and third to those in the cold water sphere, because that is the order of their impact on climate prediction for WCRP stream three.

There is considerable overlap between the requirements of WOCE and TOGA in the tropics with respect to seasonal variation (including inter-annual variability as defined above) and water mass conversion, especially in the boundary layer and equatorial undercurrent. It is recommended that the TOGA project take over all these activities in the tropics, leaving WOCE to concentrate on higher latitudes. A joint WOCE-TOGA panel should be established to ensure that all WOCE requirements are met in the tropics.

A similar arrangement might be appropriate for seasonal variation and water mass conversion in polar waters, where sea ice is important. Consideration should be given to establishing appropriate links with MIZEX (for the Arctic) and the forthcoming WCRP Antarctic sea ice project.

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CAPTIONS

Fig. 1 Seasonal variation of sea surface temperature. Sensitivity to seasonal variation in weather. Model calculation using Bunker monthly mean data.

Fig. 2 Meridional variation of the maximum penetrations of:
(1) solar heating (10 MJ/m^2y & 100 MJ/m^2y)
(2) the mixed layer in winter

Fig. 3 The seasonal variation of processes in the boundary layer shown schematically.

Fig. 4 The spring accumulation of fresh water in the upper ocean measured at one kilometre intervals along a 2.5 Mm batfish section between the Azores and Greenland Note the transition from excess of evaporation over precipitation in the south to the opposite in the North also the effects of advection at the polar front.

Fig. 5 The seasonal variation of mixed layer depth. Diurnal maximum and minimum for constant weather at a site where the net annual surface energy flux is zero.

Fig. 6 A typical temperature profile showing in which D the annual maximum depth of the mixed layer is marked by an elbow.

Fig. 7 Reid's (1982) map of D (the annual maximum depth of the mixed layer) based on oxygen measurements.

Fig. 8 Maps of D from various authors.
a. Zubov 1978
b. Robinson, Bauer & Schroeder 1979
c. Levitus 1982 (based on a temperature criterion)
d. Levitus 1982 (based on a density criterion)

Fig. 9a The meridional variation of D shown by the haloclinicity elbow in a summer hydrographic section.

Fig. 9b The individual profiles from which this section was constructed show no clear elbow marking D.

Fig. 10 Variation in the annual mean mixed layer temperature at OWS "C" (Rodewald 1983, unpublished)

Fig. 11 Seasonal mean and interannual variation in mixed layer salinity at North Atlantic OWS (Taylor & Stephens 1980)

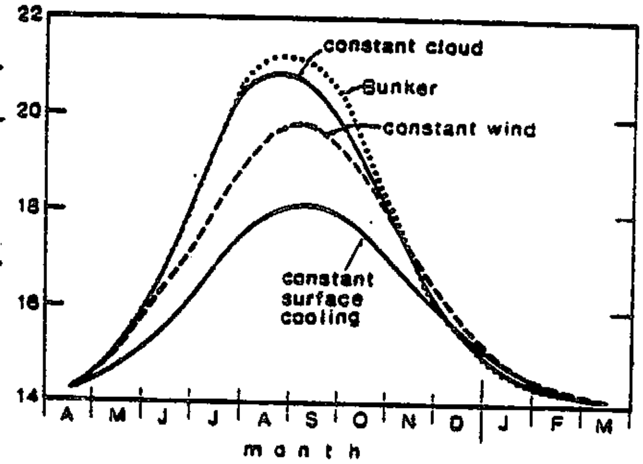
Fig. 12 The relationship between mixed layer temperature and upper ocean heat content (Gill & Turner 1976).

Fig. 13 Seasonal variation of sea ice around Antarctica.

Fig. 14 Estimates of turbulent mixing in the equatorial undercurrent based on microstructure measurements (Gibson 1983)

Fig. 15 The dependence of salt eddy diffusivity on the density ratio R, and the apparent diffusivity of R. (Schmitt 1981)

Fig. 16 Schematic illustration of cross-frontal intrusions due to thermoclinic sloping convect (Garrett 1982)



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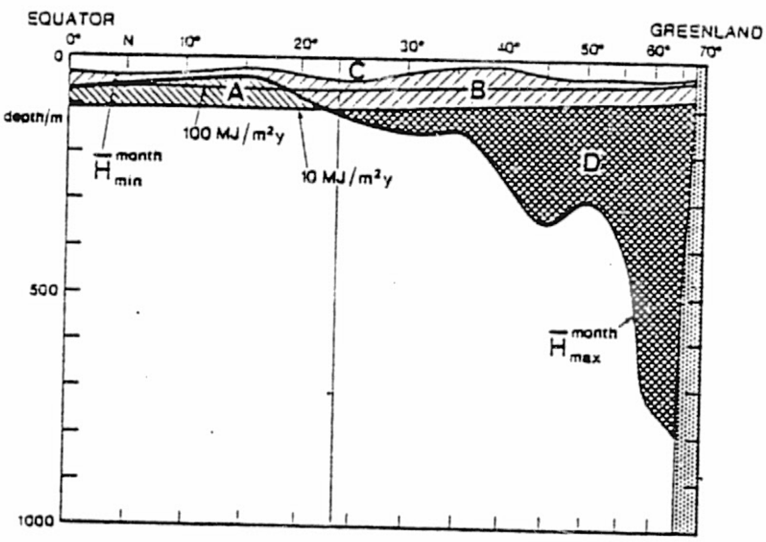

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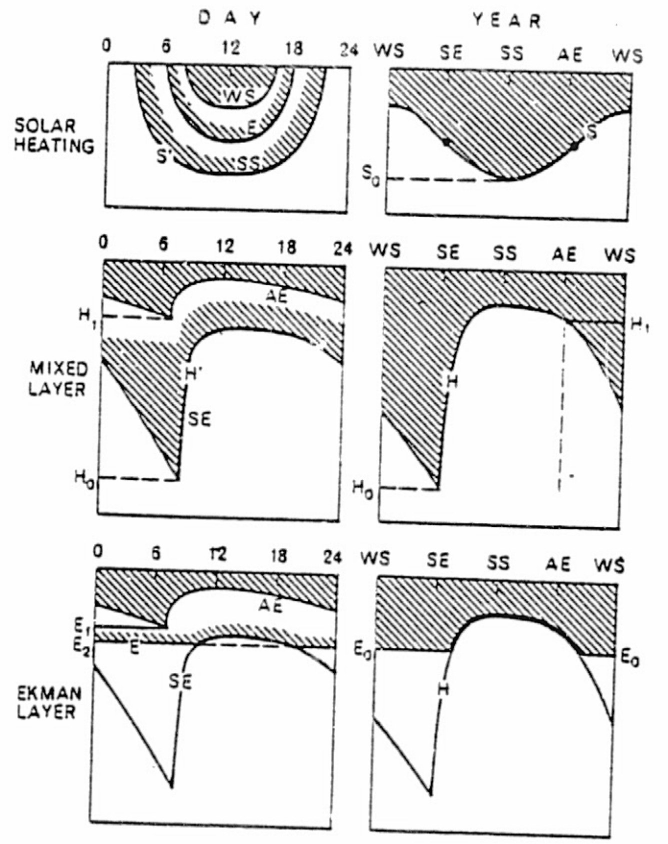

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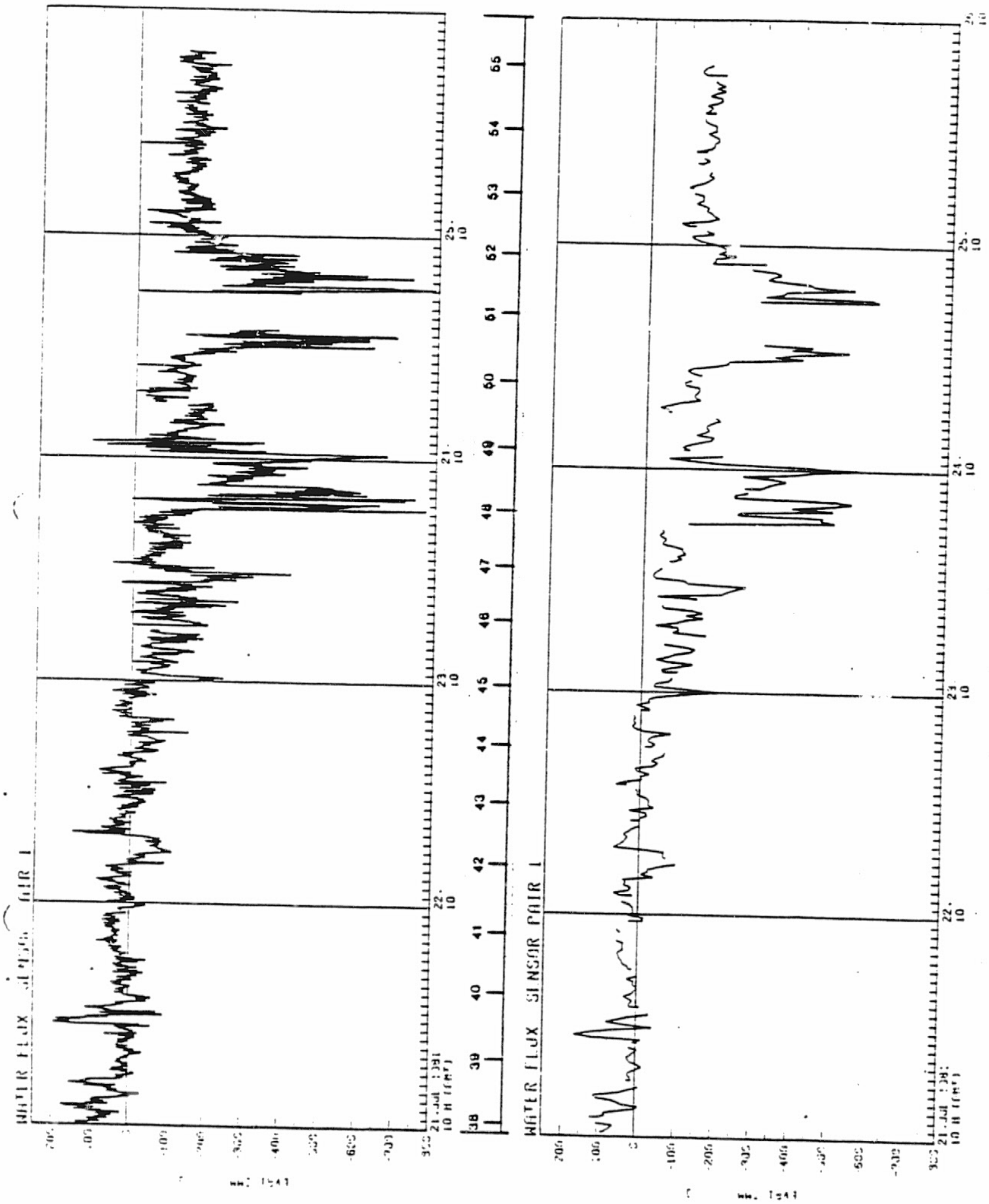

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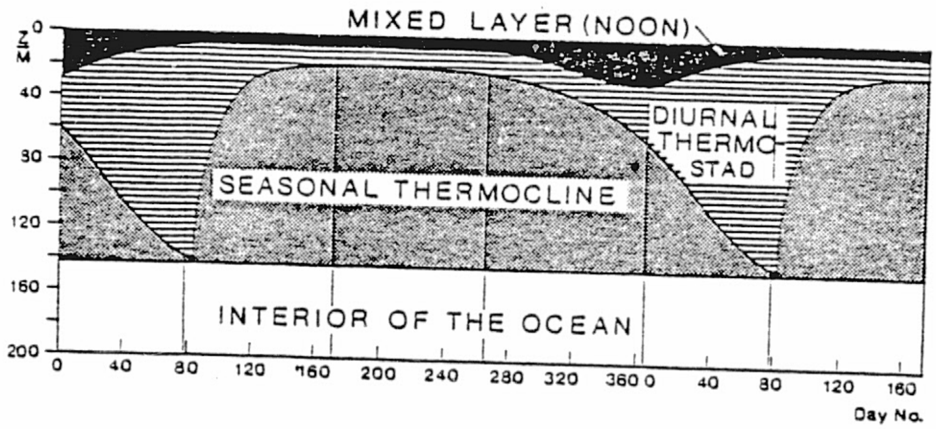

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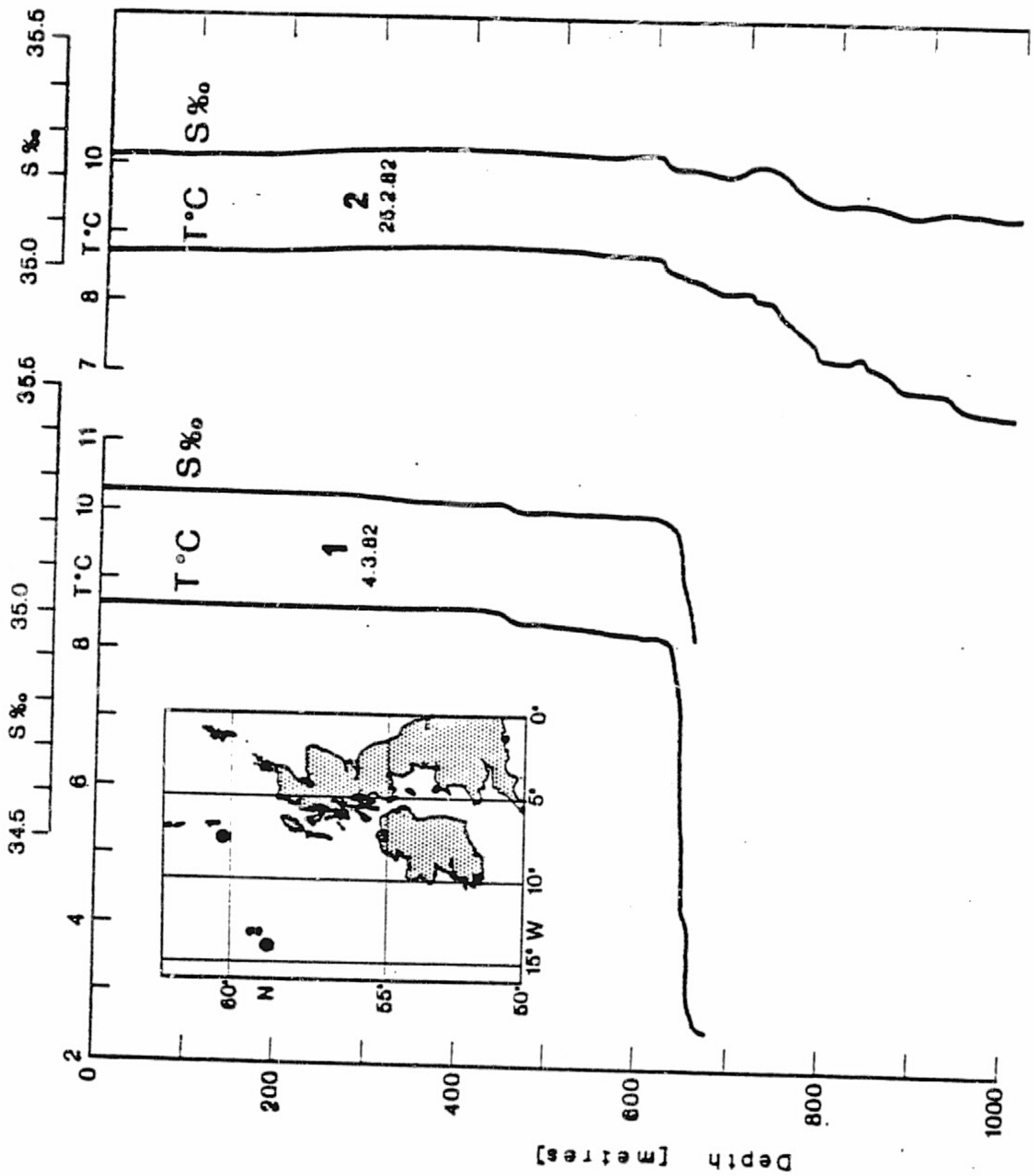

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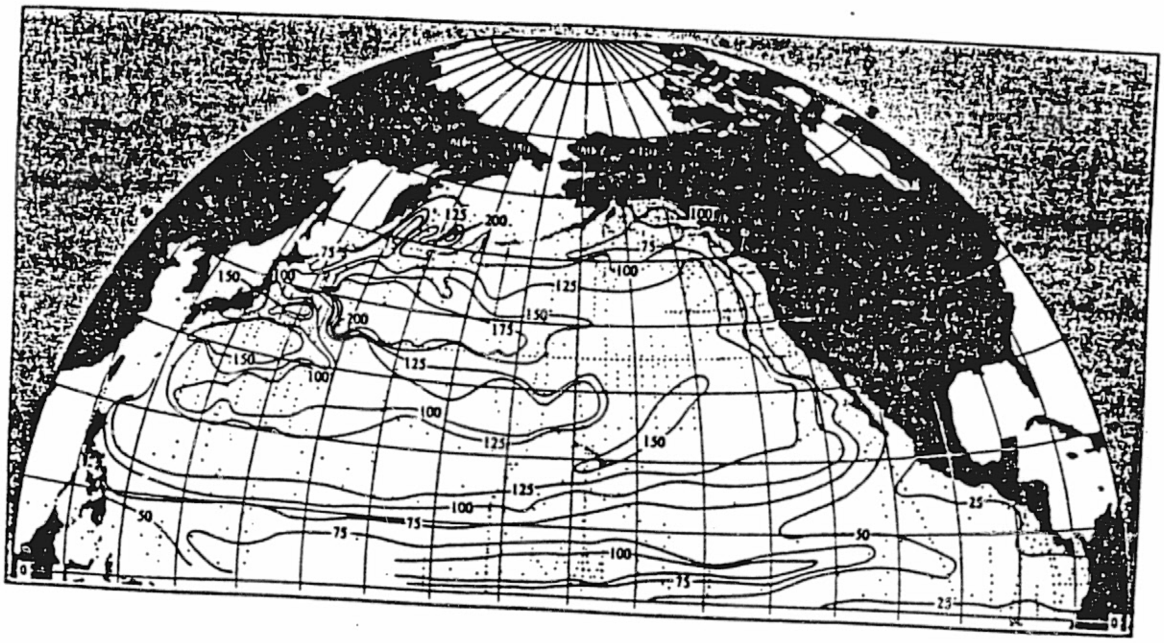

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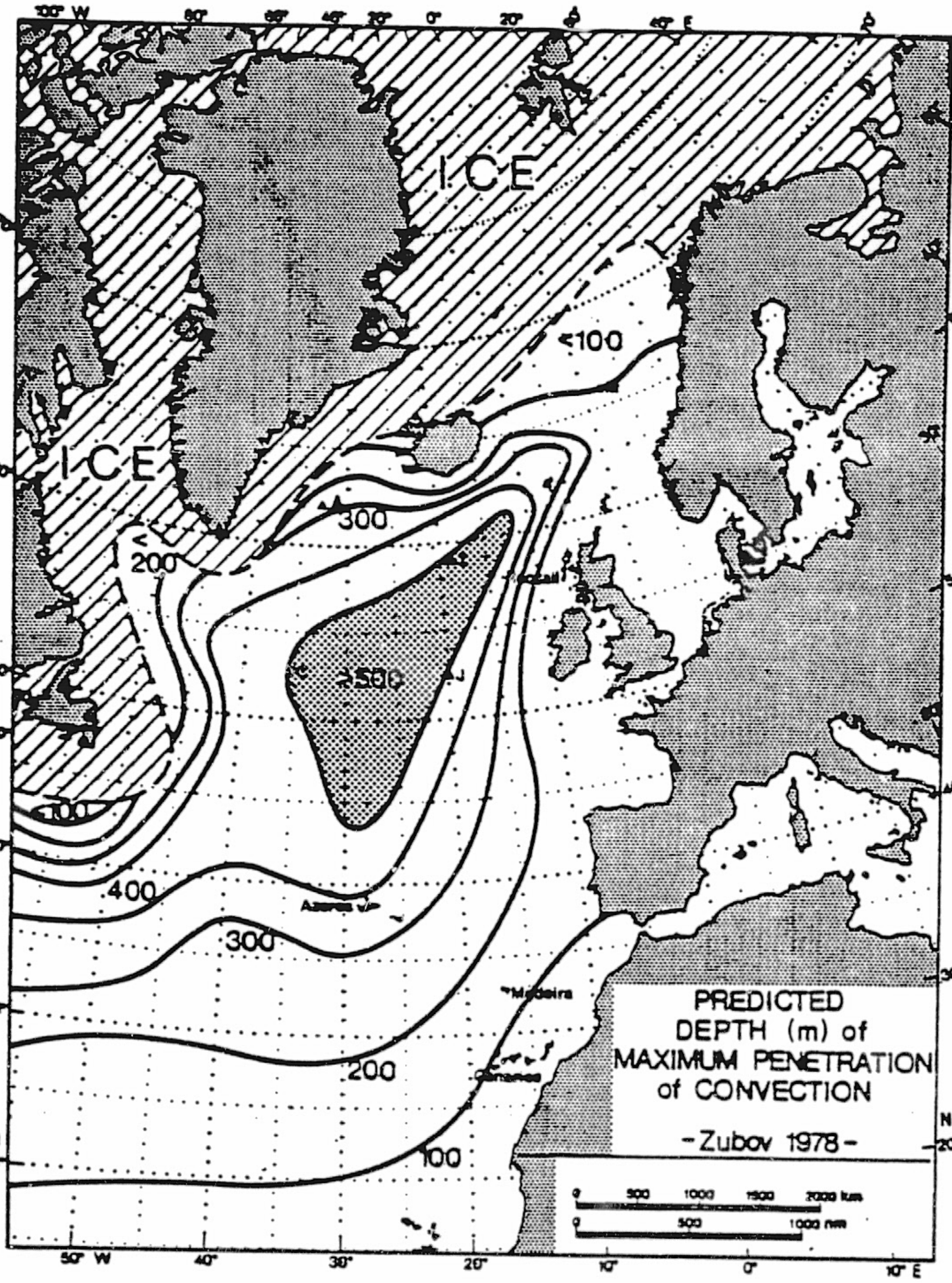

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a. Zubov 1978

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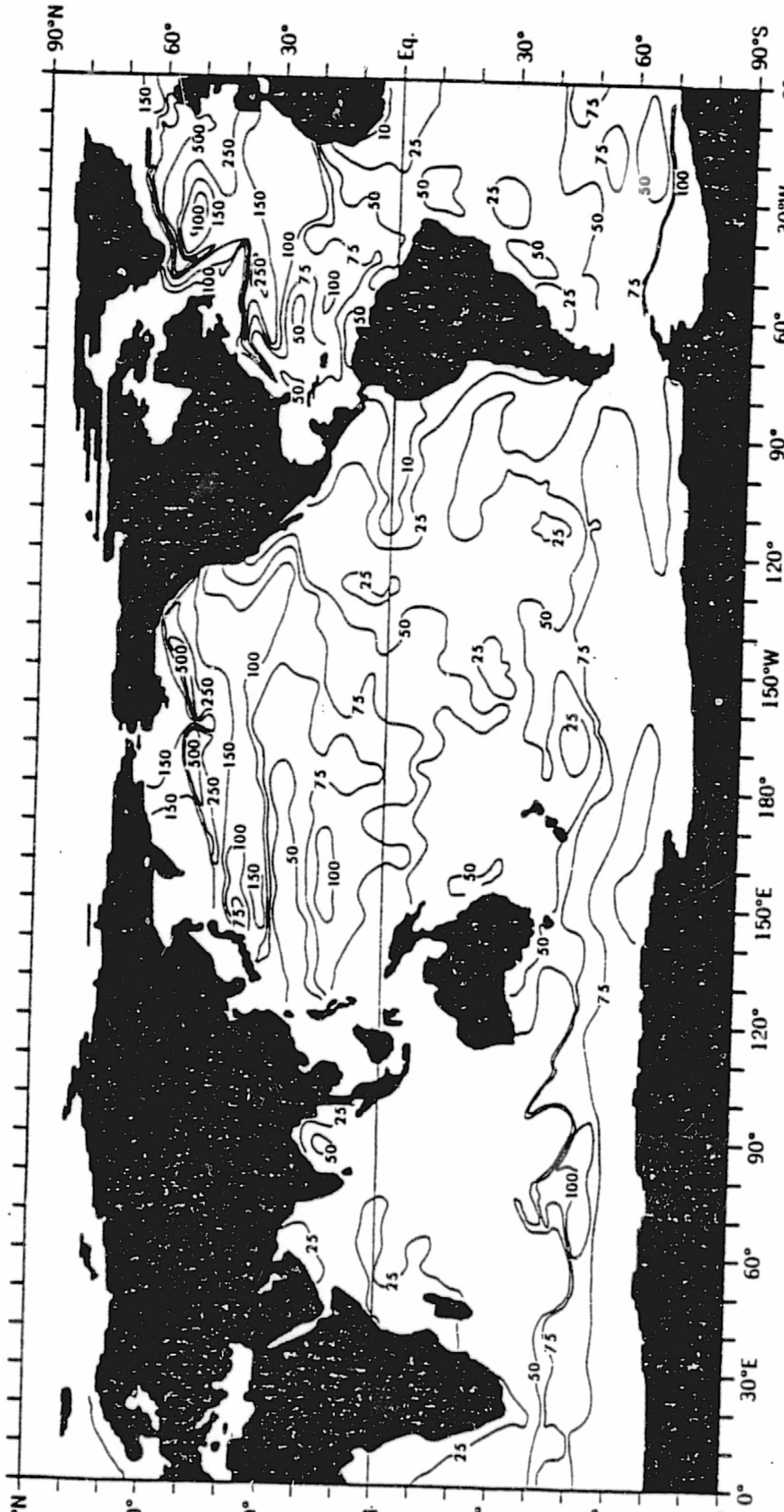
c. Levitus 1982 (based on a temperature criterion)

d. Levitus 1982 (based on a density criterion)


OBSERVED MEAN DEPTH (m) of the CONVECTIVE LAYER in MARCH
Robinson, Bauer, Schroeder 1979
1:40 000 000
Madeira
Canaries
Azores
200
300
<300
400
500
600
100
300
>300

Fig. 8 b



90°N
60°
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Eq.
30°
60°
90°S
0°
30°E
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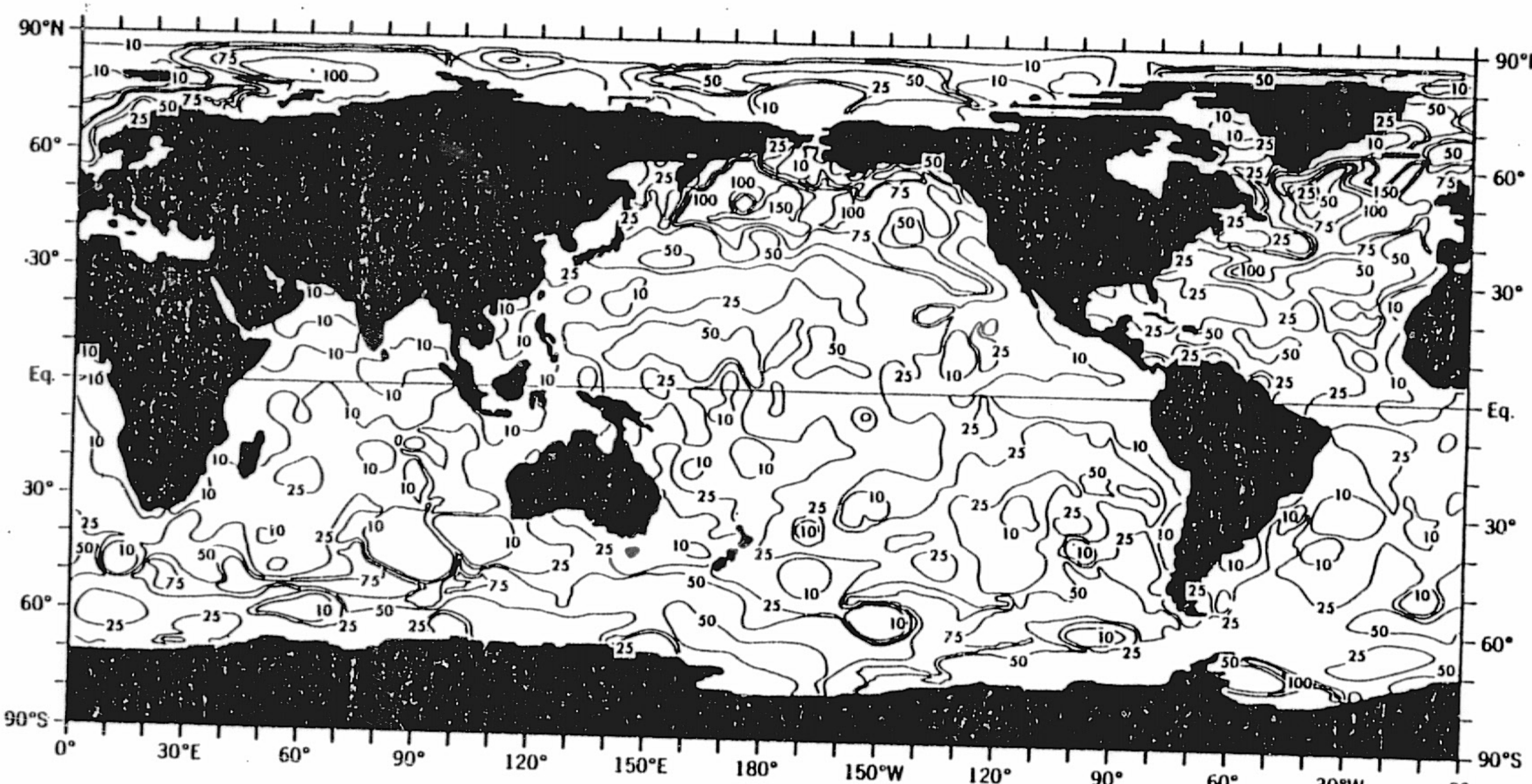
90°
120°
150°E
180°
150°W
120°
90°
60°
30°W
0°
10
25
50
75
100
150
250
500

Fig. 8 c


90°N
60°
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30°
60°
90°S
0°
30°E
60°
90°
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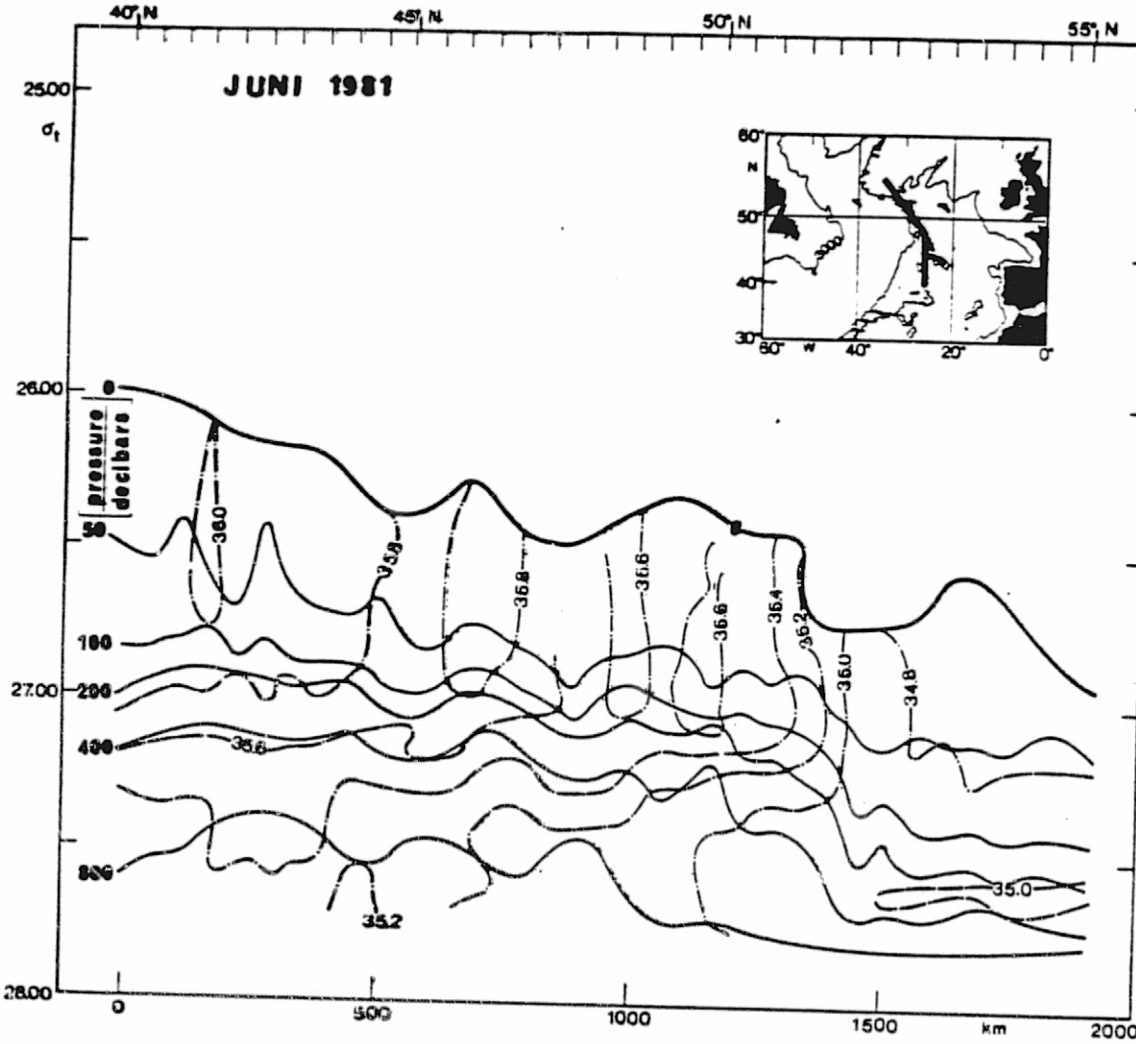
150°E
180°
150°W
120°
90°
60°
30°W
0°
90°S

Fig. 8 d



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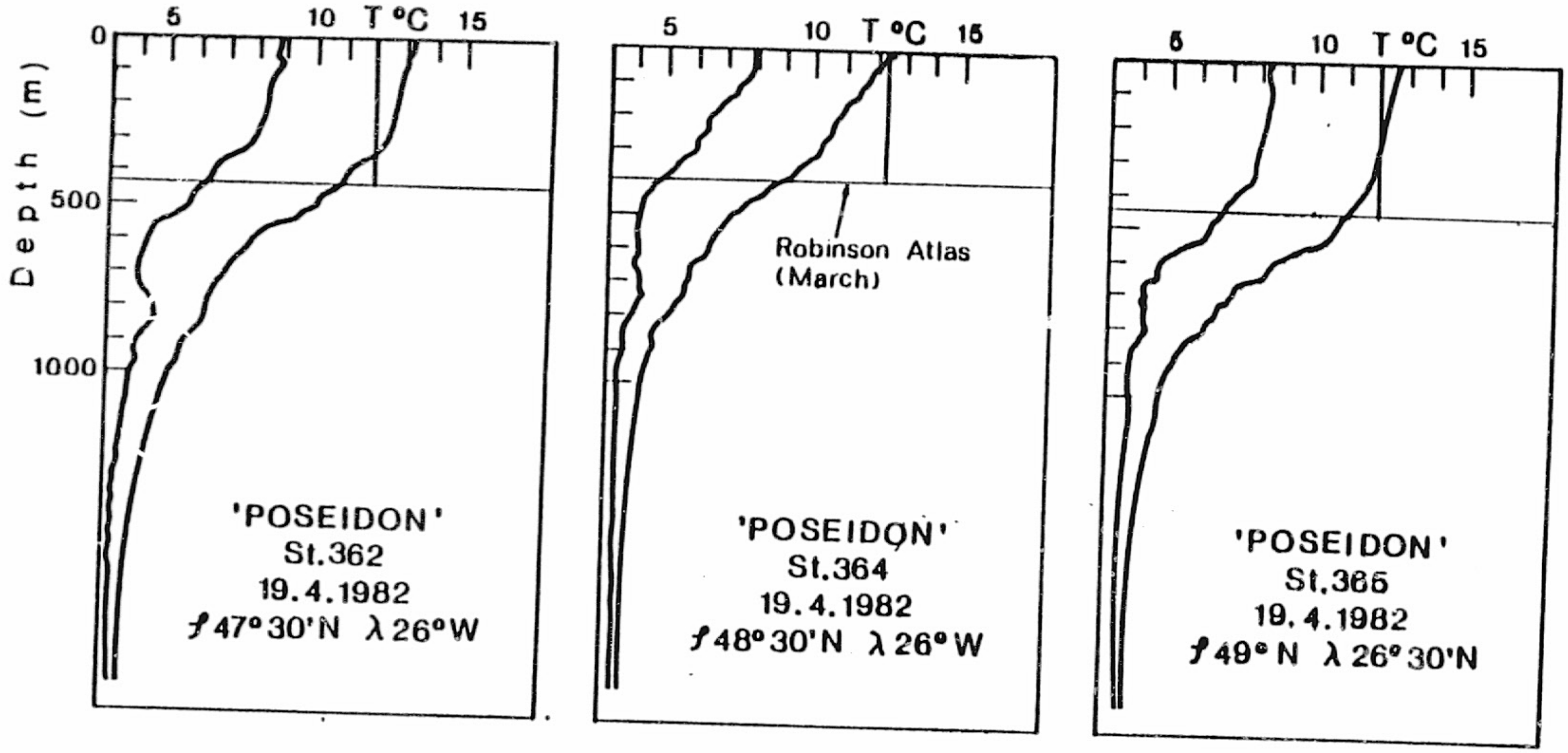

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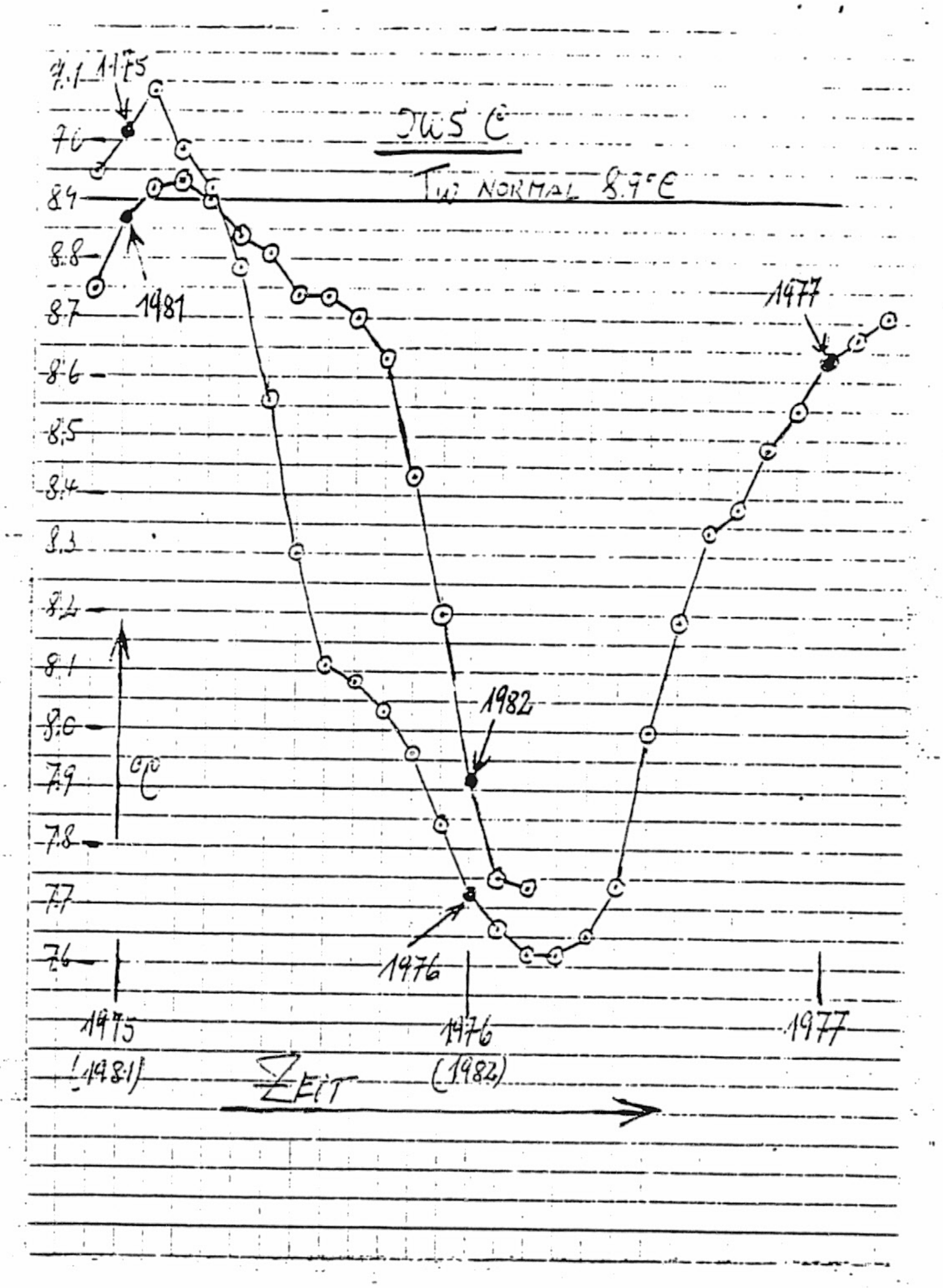

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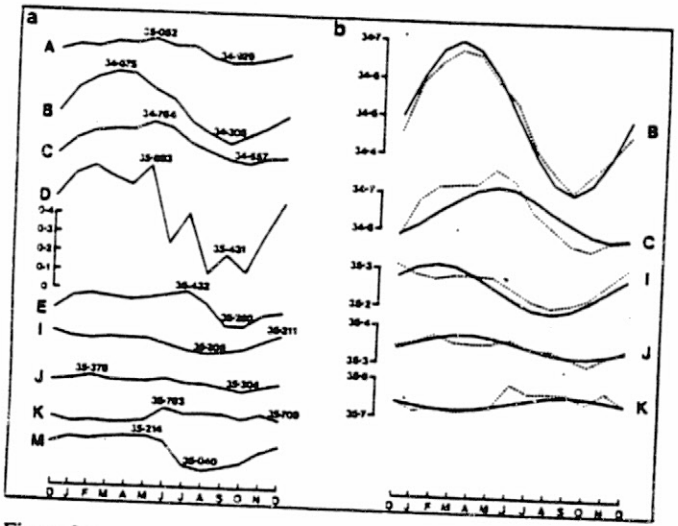


Figure 2 *a*

Long-term monthly mean sea surface salinities and temperatures at the nine Ocean Weather Stations.

Figure 2 *b*

Predicted (unbroken line) and observed (dotted line) seasonal salinity variations at (B) Bravo, (C) Charlie, (I) India, (J) Juliett and (K) Kilo.

a

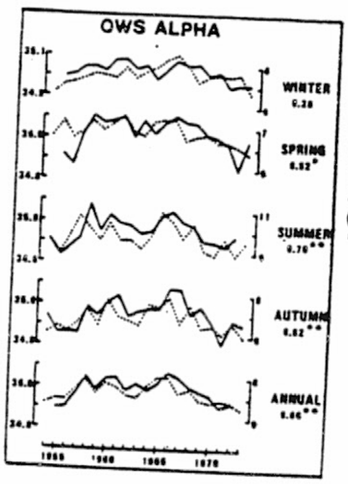


Figure 4

Station Alpha: interannual variation of salinity (unbroken line) and temperature (dotted line) by season. The correlation coefficients between the salinity and temperature time-series are included ($P < 0.05$; ** $P < 0.01$; *** $P < 0.001$).*

b

Fig. 11 Seasonal mean and interannual variation in mixed layer salinity at North Atlantic OWS (Taylor & Stephens 1980)

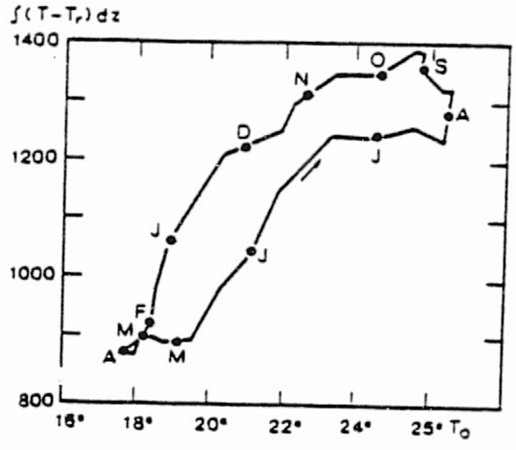


Fig. 12 The relationship between mixed layer temperature and upper ocean heat content (Gill & Turner 1976).

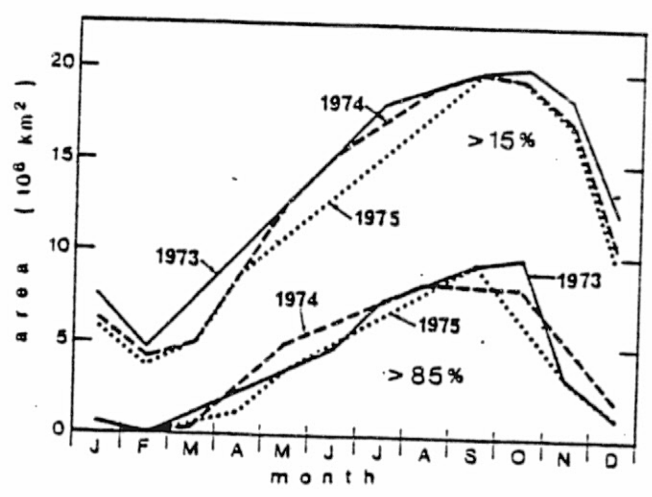


Fig. 13 Seasonal variation of sea ice around Antarctica.

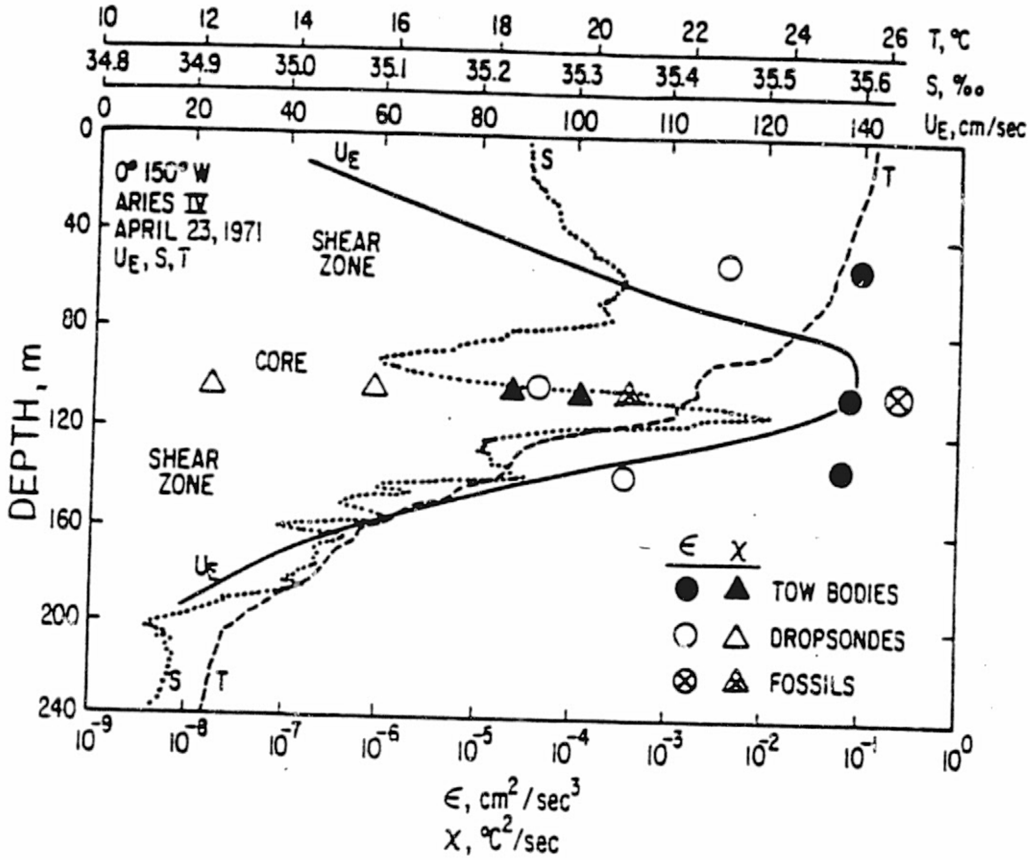


Fig. 14 Estimates of turbulent mixing in the equatorial undercurrent based on microstructure measurements (Gibson 1983)

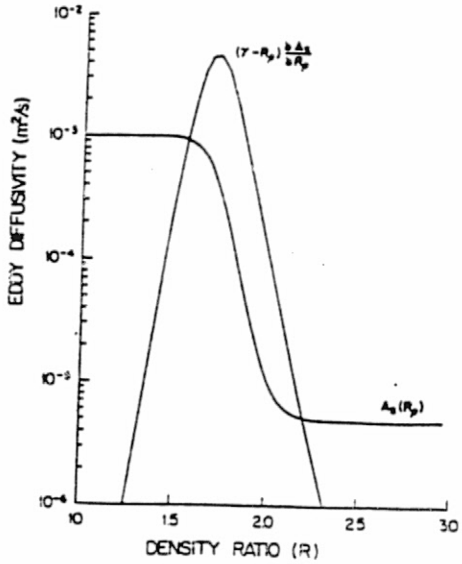


FIG. 6. The dependence of salt eddy diffusivity on R_ρ for the model calculations. Also shown is the apparent diffusivity for R_ρ, $(\gamma - R_\rho)\partial A_S/\partial R_\rho$.

Fig. 15 The dependence of salt eddy diffusivity on the density ratio R, and the apparent diffusivity of R. (Schmitt 1981)

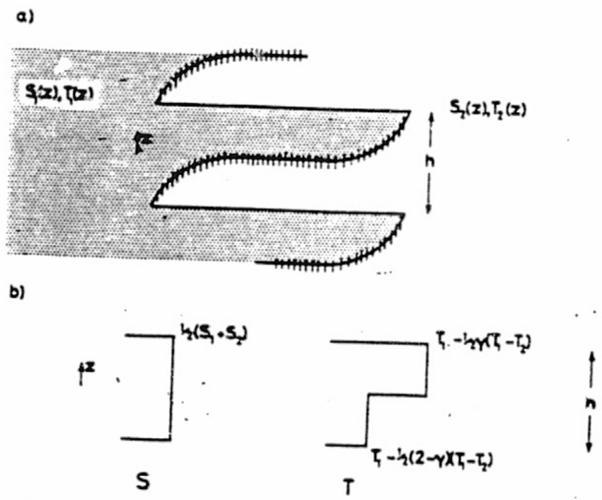


FIG. 2. Schematic (a) of double-diffusive intrusions across a thermohaline front, with salt fingers from the saltier waters (shaded) to the less salty (clear); and (b) the assumed profiles of S, T, in the run-down state, denoting $S_1(0) = S_1$, etc.

Fig. 16 Schematic illustration of cross-frontal intrusions due to thermoclinic sloping convect (Garrett 1982)

Appendix

Deep-Sea Research, 1973, Vol. 20, pp. 141 to 177. Pergamon Press. Printed in Great Britain.

The theory of the seasonal variability in the ocean

A. E. GILL* and P. P. NIILER†

(*Received* 31 *May* 1972; *in revised form* 30 *August* 1972; *accepted* 5 *September* 1972)

Abstract—The theory of large-scale seasonal variations of temperature, salinity, sea-level, velocity, etc. in the ocean is considered herein. Regions near the boundaries and within 15° of the equator are excluded. It is found that, for the scales considered, the heat input is mainly stored locally and horizontal advection by the mean flow is *not* particularly important. Effects of vertical advection and of seasonal changes of horizontal advection in the Ekman layers are calculated for 5° squares in the North Atlantic and North Pacific and also found to be relatively minor. Observational evidence is discussed.

Contributions to sea-level changes are calculated for each season for 5° squares in the North Atlantic and North Pacific. First, there is a response to changes in atmospheric pressure which is particularly important (a few cm) at high latitudes. This response is, however, dynamically uninteresting as temperature, velocity, etc. are hardly changed. Second, there is the *barotropic* response to changes in the wind stress. This produces small changes (a few mm) in sea-level on the scales considered, and corresponding changes of bottom pressure. Velocity amplitudes increase to about 3 mm sec^{-1} in the west, corresponding to horizontal displacements of 15 km. Although these values seem small, the associated transports, spread over 20° of latitude, are considerable (30×10^6 m^3 sec^{-1}). However, since topographic effects are so important for barotropic motions, the large transports are confined to deep water regions and would not, for instance, be expected to contribute much to seasonal changes of transport through the Florida Straits. Because the changes in bottom pressure are small, the sea-level changes are approximately *isostatic*, as concluded from observational studies. On the other hand, there is reason to believe that large barotropic changes will be found in regions of closed contours of H cosec ϕ, where H is the depth and ϕ the latitude.

Thirdly, there are *steric* changes in sea-level. The major change (a few cm) is produced by expansion and contraction of the water column above the seasonal thermocline due to changing fluxes of heat and water across the surface. Currents of a few mm sec^{-1} are produced above the seasonal thermocline by the changes in density field, but the transport over 20° latitude is under 10^6 m^3 sec^{-1}. There are other, less important, steric changes (usually less than 10%) due to movements induced by the changing wind stress, and these effects have been explicitly calculated by season for the North Atlantic and North Pacific. The changes involve (1) changes in the *mixed layer* due to convergences of heat and salt produced by Ekman fluxes, and (2) changes produced by Ekman pumping which displace the main thermocline up or down by a few metres. The latter change represents the *baroclinic* response which increases in strength towards the equator, although even at 15° (where the analysis breaks down), currents are only about 1 mm sec^{-1}. The baroclinic response dominates the barotropic response at low latitudes, while the reverse is true at high latitudes. The two responses are comparable at 30°. Fourthly, there is a tidal component, *Sa* (DOODSON, 1921; WUNSCH, 1967) with a period of 1 year and an amplitude of a few mm which will not be discussed herein.

Large-scale *anomalies* in the surface temperature of the ocean are discussed qualitatively, although these are not seasonal changes. It is suggested that the main changes are simply due to changes in the heat flux through the surface, and changes in the convergence of heat in the Ekman layer, the latter becoming significant at low latitudes. Simple ways of modelling the changes can be employed for regions not too close to the equator.

A World Ocean Circulation Experiment

Prepared for the "Workshop on Global Observation and Understanding of the General Circulation of the Oceans"

National Academy of Sciences
Woods Hole Study Center
Woods Hole, Massachusetts

August 8-12, 1983

Carl I. Wunsch
Massachusetts Institute of Technology

Following is a very rough, and obviously incomplete, draft of a description of a possible global ocean circulation experiment. I have combined elements of the Tokyo report, with Appendix III of the CCCO Meeting Report Number 4, along with purely personal ideas, in an attempt to show the scope of such a program. If we are to proceed with something like WOCE, then a much improved, complete version of a document such as this will have to be prepared.

For the present time, I emphasize the rough draft nature of what follows, and the fact that it has no official standing whatsoever.

Carl I. Wunsch

PRELIMINARY DRAFT 16 JUNE 1983

A World Ocean Circulation Experiment

1. Introductory Summary

The past decades have seen great progress made toward understanding of the ocean circulation. Much of this progress has occurred under the impact of technological innovation. In the decades following World War II the oceanographic community has developed a good zero order picture of the major physical and dynamical elements of the fluid flow. Geophysical fluid dynamics has provided a theoretical framework into which one could place observations obtained from a long list of novel instrumentation made available under the electronics revolution. The computer has made possible both the handling of the large data quantities generated by the new instruments, and the creation of numerical models of entire ocean basins.

As a result of these major advances, it is proposed that we are now ready to tackle what seem' to be the fundamental issues of understanding the oceanic general circulation. As recently as 10 years ago, too little was known and instruments were too crude to even contemplate the program described here. But as a result of several parallel developments, we argue that the time has come to mount a truly global experiment to observe and understand the ocean. We mention only three: 1) The regional programs of the past decade, which took advantage of the newly developed instrumentation for time series of observations almost everywhere in the ocean, have made clear the problem of sampling the ocean. Programs such as MODE-1, CUEA, ISOS, INDEX, ... showed that the ocean is time variable on all measured temporal scales and is energetically dominated by comparatively small spatial scales. The result is an understanding of the magnitude of the problem of adequately observing the ocean, along with considerable knowledge of the sampling requirements that must be met.

2) The continued development of new technologies for observing the ocean. Here we would emphasize newly emergent techniques (described at greater length below) which make it possible to observe the ocean on the global scale, while at the same time meeting the sampling requirements newly raised in the recent past.

3) A general understanding that because the ocean is a global fluid, it must be observed on global scales and that societal concerns about the ocean, some of which have become urgent, can only be addressed globally. (We refer to the report by the National Academy of Sciences, 1966, providing the rationale for what became of the Global Weather Experiment as an analogous situation in meteorology).

The idea of a World Ocean Circulation Experiment thus represents a confluence of these three trends: the recognition, as a result of vast scientific and technical progress of 40+ years of the great difficulty of observing the ocean; that this same progress has brought to hand the possibility for a solution to the observational problem; and a renewed sense that understanding of the ocean is of considerable practical importance.

What follows is an outline of a World Ocean Circulation Experiment (WOCE). A number of goals of varying degree of difficulty of achievement are listed. Some approaches to these goals are described, as are a number of <u>potential</u> elements of the program. The general theme of the program is that present understanding of the ocean circulation is predominantly limited today by a lack of observations. Much of WOCE (or strictly speaking, the problem of designing WOCE) is directed at remedying the paucity of observations. But the observational program is constructed around detailed theoretical knowledge of how the ocean circulation probably works. Deployment and ultimate use of the WOCE observation systems and data is dependent upon theoretical and modeling efforts; observational and theoretical elements are inseparably linked.

2. Primary Goals:

The overall goal of WOCE is to greatly improve understanding of the general circulation of the world oceans. Within this general goal are a large number of more specific goals among which we identify:

1) Quantitative determination of the long-term average oceanic fluxes of heat, fresh water, of bio-chemical tracers, and of man-induced tracers such as fluorocarbons and tritium.

2) Determination of the annual and interannual fluctuatuions in these and other fluxes.

3) Determination of the primary fields of air-sea coupling, including wind-stress, air-sea heat and moisture fluxes, their multi-year means and their annual and interannual variability.

4) Understanding of the oceanic responses to imposed changes in air-sea forcing functions.

5) Determination of the oceanic rate of uptake of atmospheric gases (CO_2, fluorocarbons).

6) Determination of the space and time scales, and energy levels of oceanic mesoscale variability.

7) Determination of the nature and importance of oceanic variability on middle (large than mesoscale) and basin scales.

8) Improved understanding of the nature and rates of oceanic mixing including convective overturning (water mass conversion rates).

9) The production of quantitative tests of the validity of ideas concerning the primary dynamical balances governing large scale oceanic motions.

10) Design of an effective and economical system for measuring and monitoring the ocean on climatological scales in the period following WOCE.

These goals simultaneously served a number of important concerns:

(1) One of the major goals of the World Climate Research Program (WCRP) is to understand the role of the ocean in determining our present global climate, its role in fluctuations of the climate state, the rate at which the ocean will absorb the current CO_2 transient, and the way in which the ocean will respond to changes in atmospheric circulation under the CO_2 induced warming.

(2) The general problems of the physical oceanographer in trying to understand what the general circulation of the ocean is and what are its dominant physical mechanisms.

(3) The need to understand the general circulation of the ocean in adequate detail to predict the fate of wastes in the ocean over very long time-scales with particular concern for potential high level radioactive wastes.

(4) Important issues concerning the biological cycles in the ocean are probably unaddressable until there is vastly greater understanding of the physical and chemical environment in which these cycles occur.

The design of WOCE is based upon the belief that these and other important problems are all best addressed through a comprehensive, multi-year, program to observe the global ocean circulation and its changes. Such a program is proposed for the late 1980's to early 1990's because of the confluence of several factors:

1) Much increased understanding, from research of the past several decades, of the nature of the ocean circulation, and reasonably complete understanding of the temporal and spatial sampling issues.

2) The development at many different institutions in many countries of instrumentation capable of long time series measurements at arbitrary depths in the world oceans.

3) The continuing development of numerical models of basin and global scale ocean circulation with resolution adequate for all energetic scales of motion.

4) Progress in the development of methods for rapid and cheap determination of chemical tracers.

5) Developments in satellite technology providing for the first time, the possibility of global scale observations of both the ocean (altimetry) and of the primary forcing function (scatterometers).

6) The realization that most of the important problems mentioned above are unanswerable without such a program.

3. <u>Observational Foci:</u>

1) The fields of surface forcing as determined by

a) Satellite scatterometers (anticipated spacecraft are the European Space Agency's ERS-1 and the U.S. Navy's NROSS) for wind stress.

b) Conventional meteorological analyses for wind stress and heat and moisture fluxes based upon improved atmospheric modelling efforts and an upgraded surface observational system using drifters in otherwise uncovered regions.

2) The field of oceanic surface topography (the oceanic surface pressure boundary condition) determined by

a) Satellite altimeters (anticipated spacecraft are ERS-1 and the U.S. NASA Topex Mission)

b) Tide gauge network on coasts and oceanic islands

3) Global hydrography from ships usng both highest quality CTD observations plus developing free fall instruments from "research ships of opportunity" to provide both a climatological T-S data base plus continent to continent, top-to-bottom section for use with dynamic method and ancillary observations and models.

4) Global measurements from ships (probably combined with hydrographic measurements) of "light" i.e. small sample, chemical tracers, including tritium, fluorocarbons, oxygen, nutrients... These observations, along with the hydrography provide the primary observables representing the integrated (i.e. time-average) behavior of the ocean circulation.

5) Global, but sparse observations of "heavy", i.e. large water sample, chemical tracers such as carbon-14 as part of the determination of time-integrated oceanic behavior.

6) Global, but sparse, determination of the dominant source-sink terms of the bio-geochemical tracers involving transfers to and from the seafloor and continental margins by sediment traps etc. for purposes of using the tracers.

7) Direct observations of the flow fields over large areas for long time periods by "clouds" of neutrally buoyant floats and surface drifters, both communicating directly with satellites.

8) Measurements of large areal averages for long times, of heat content, velocity and vorticity and potential vorticity through moored acoustic tomography arrays.

9) Ship of opportunity programs involving near-surface velocity measurements through acoustic backscatter and of direct near surface density field through xbt's.

10) A few, special, regional foci involving ships, conventional moored current meters, floats, temperature recorders, tide gauges and other in situ instrumentation. These regional foci would be chosen as those places where local knowledge had direct and immediate impact upon knowledge of the larger circulation scales. Candidate regions are the overflows of the northern North Atlantic, convecting regions of the North Atlantic, the flux through the Florida Straits, or the flux of the Kuroshio and other western boundary currents, the Drake Passage flux, etc.

4. Modelling Foci

Several different types of modelling effort will be important to WOCE.

By the time of the major field effort we expect to have available eddy resolving basin scale general circulation models with good parameterization of non-adiabatic processes. We already have global scale general circulation models with thermodynamic components. Both types of models, and anticipated hybrids, would be used with the WOCE global forcing functions (winds and surface pressure boundary conditions) and with the interior observations of velocity, temperature, vorticity etc. "assimilated" into the models for analysis.

A special sub-set of these models will be those used to study chemical tracers. As already noted, the tracer observations are directed at the near-zero frequency behavior of the ocean. As with the other observables, but with different emphases and techniques, the observations will be used with models in various modes and methods ("assimilation", "optimal estimation", "inversion", etc.) to make inferences about oceanic physics.

We expect that analytical and semi-analytical modelling will become very much more active than in the past, under the impact of both the observations themselves and from inferences from the numerical models used in conjunction with the observations.

For reasons (technical) discussed below, WOCE will include major numerical tide and gravity field modelling efforts.

5. Timetable

The major field effort in WOCE will be timed to coincide with the flight of at least one altimetric and one scatterometer mission. At present, the period envisioned is 5 years beginning in early 1989. A 5-year period for intensive observations has been chosen for a number of reasons. It is a period of sufficient duration to yield good estimates of the time average circulation, several realizations of the annual cycle, and of the interannual variability. It is also consistent with the maximum expected lifetime of spacecraft missions (e.g. five years for Topex) and is of adequate duration to permit global surveys by ship without overburdening the research institutions in any given year.

A number of WOCE efforts must however take place prior to the intensive field phase. Much of the technology described above is still in the development phase; it must be tested and made sufficiently reliable in the years between now and 1989. WOCE will need numerical models of types that do not now exist, but which are under development—developments that need to be accelerated. Many other programs (e.g. the U.S. Transient Tracer Experiment) are already occurring and will provide "reconnaissance data" for the detailed planning of the intensive field phase.

6. Scope

The emphasis within WOCE is on the global scale. As with any fluid system, complete understanding can only come when all energy dominant components are measured and the system is not energetically open. The global scale is beyond the reach of existing consortia of oceanographers and access

to it is the only justification for creating a large international scientific program. Existing organizations are fully capable of mounting regional studies. The global scale requires a higher level of collaboration, and the use of resources, such as satellites, which involve governmental bodies at high levels. WOCE is envisioned as supplementing, not replacing, existing arrangements and programs. Inevitably, there will have to be some redirection of effort during the main field program of WOCE (financial and human resources are both limited), but regional and process oriented oceanography will remain important elements of the science.

7. Elaboration on Observational Foci

7.1 Surface Forcing--Winds

To the extent that the ocean circulation is taken as decoupled from the atmosphere (i.e. ignoring the feedbacks of the ocean on the atmosphere) the circulation is forced by winds and by air sea transfers of heat and freshwater. Over most of the ocean it is not the wind stress itself but rather its curl that is the dynamically crucial quantity; the tropical regions are an exception. At the present time, in order to determine the forcing by wind, oceanographers are forced to rely primarily upon crude shipboard observations. These observations are mostly confined to major shipping lanes, are of uncertain quality and are subject to a bias because ships tend to avoid regions of high winds. Thus extant estimates of the wind field over the sea are crude climatological averages taken over long times and over large distances. It is impossible to make valid estimates of what the stress is in any particular month of any particular year - a situation that becomes intolerable when one seeks to understand the state of the ocean circulation at a specific time.

The situation for estimating thermodynamic forces is even worse. Evaporation and the transfers of latent and sensible heat between the ocean and atmosphere are normally based upon so-called bulk aerodynamic formulas which involve empirical drag coefficients, the wind speed to some high power, and often such ill-determined quantities such as air-sea temperature differences. Some critical quantities involve estimating cloud cover--notoriously difficult to observe (Charnock et al., 1982 discuss some of the problems). But improvements in the wind field estimates would go far to reduce the uncertainties in many of the most important transfer processes.

Thus obtaining estimates of the wind stress over the sea so that the mean and variability of the major forces acting upon the ocean may be determined must be a major goal of any global circulation experiment. The stress must be determined to a fractional precision better than other elements in the system to minimize uncertainties in any comparison of model performance with observations. The need is for seasonal evolution of time averaged stress for a few typical years; the spatial resolution should be appropriate to the structure of this time averaged stress in specific regions and is typically in the range of a few hundred to one thousand kilometers. The strategy for meeting this requirement involves a satellite-based scatterometer in conjunction with a programme for systems in situ (both wind speed and surface pressure distributions). Given an instrument relatively free of directional ambiguities, and excepting unsuspected sources of systematic error which may become apparent as the physics of the relationship between backscatter cross section and surface wind become better understood, the residual uncertainty after averaging large numbers of observations should be less than ±10% or 0.5 m/s in the vector wind, whichever is larger. Such overall system accuracy would yield a major improvement over current knowledge. Important next steps

are the definition of the observing system's intercomparison programme, and analysis of the potential impact of biases introduced in the resolution of directional ambiguities and by undetected rainfall.

7.2 Surface Forcing--Heat Flux

The sea surface temperature is a significant variable in the context of modelling ocean/atmosphere interactions only if accompanied by knowledge of the net surface heat flux and its sensitivity to sea surface temperature changes on various scales. The field of net heat flux coupled with this sensitivity constitutes another fundamental forcing for the general circulation of the ocean. This forcing is poorly known, yet it is central to the overall understanding of ocean atmosphere interactions.

Some of the components of the heat flux can be determined better than others. A particular problem concerns the evaporation rate, one of the largest components. Even under favorable observational circumstances, there are significant uncertainties in current methods of determining this on a large scale and the prospects for significant improvement in present knowledge of the evaporation rate over the globe by present techniques are poor. It is a high priority matter to develop observational techniques in order to make better estimation of net evaporation possible.

7.3 Surface Forcing--Moisture Flux

Because it cumulatively affects the salinities of the upper layers, and hence preconditions water masses for convective overturning, net moisture flux (evaporation minus precipitation) is also a fundamental forcing and the evaporation rate is a major uncertainty in the net surface heat flux. Unfortunately, over oceanic areas the cumulative precipitation is probably

even less well known than the evaporation and the net difference is uncertain to perhaps a factor of two over wide areas.

The designation of ±10 W/m^2 as the desired accuracy for surface heat flux (see the Cage Report, WCP-22) if applied to the component associated with atmospheric moisture divergence alone, implies a net moisture flux of ±15 cm/y. If attainable, this would be a very useful measurement. The prospects for dramatic improvement by traditional approaches are poor and a spectrum of radical new strategies needs to be explored. The need is great.

One possibility is to infer the net surface moisture flux directly from the atmospheric moisture flux divergence over large areas, using an atmospheric assimilation model strengthened by direct observations of vertically integrated precipitable water (which have been obtained on a pilot basis from a satellite-based microwave radiometer), coupled with the maximum available information about the low-level wind field (e.g., from cloud winds). This approach could be tried using existing data archives but does not appear to have been given major attention.

7.4 The Field of Surface Topography

In seeking critical observables for understanding the ocean on a global scale, one is faced with many important physical constraints. The size of the ocean means that the physical variable must be observed globally; whatever one chooses to observe must have some directly computable relationship to flow dynamics; finally the variable should not be overly sensitive to unobservable local physics. The global scope necessarily drives one toward satellite techniques, but the ocean is opaque to electromagnetic radiation, and whatever one could hope to measure will be a surface variable. The sea surface has many measureable physical properties—roughness, color, dielectric constant, temperature, elevation. Roughness can be measured in a variety of ways

(synthetic aperture radars, scatterometers, altimeters). Suitably interpreted, roughness provides a measure of the stress acting on the ocean by the atmosphere and is the basis for the use of scatterometers for inferring the wind stress. Thus (apart from isolated problems like internal wave modulation of the surface) roughness is primarily a measure of how the ocean is being forced, not how it is responding.

Infrared temperatures (IR) of the sea surface have been measured longer than any other space observable. With suitable accuracies these are very valuable measurements both for providing boundary conditions upon atmospheric models where the ocean temperature is a boundary condition, and for showing the enormous complexity of near-surface physics. But as a tool for understanding the ocean, IR (and the more recent microwave measurements) do not have great promise for fulfilling the need of those examining the circulation, although the measurements are important--on large scales--for understanding air/sea heat transfers. The reasons are several: a) the temperature measured is that of the upper fraction of a millimeter of the sea, b) this temperature seems to be not easily related to the circulation even a few meters below, because much microscale physics operates at the sea surface not easily related to the larger scale, deep flows.

Ocean color is measured down to an optical depth of several meters and is a reflection not only of the large scale circulation but also of surface temperature, biological productivity and sediment transport. It contains all the complexity of the surface temperature measurement as well as the added complications of biology.

Surface elevation emerges as a uniquely useful variable in this context; it is described at some length by the Topex Science Working Group, 1981.

There are a number of reasons for this: a) Surface elevation is a dynamical variable and can be treated directly as a boundary condition on the large scale flow (along with the wind stress or windstress curl). b) Near surface flow processes, often of great complexity, do not show up to first order as pressure, i.e. surface elevation changes--surface elevations primarily reflect the largest scale (quasi-geostrophic) flows below the top 50 to 100 m of the ocean. c) Altimeters are not limited by the presence of clouds, d) (related to (b)) the slopes of the sea surface topography immediately provide the geostrophic velocity which is directly related to solving the oceanographer's infamous level of no motion problem.

Experience with altimetric satellites (and with existing arrays of island tide gauges) support the idea that altimetry is capable of providing a global determination of variability on all time and space scales, and of providing the time average sea surface topography (the pressure boundary condition) on large spatial scales.

7.5 Chemical Tracers

Chemical tracers are extremely diverse; they come in most combinations of stable/unstable, transient/steady-state, conserved/non-conserved, dynamically active/passive. The use of salt and temperature as dynamically active, stable, steady, sometimes conservative, tracers goes back for more than 150 years. At the opposite extreme fluorocarbons are of very recent use and represent a transient, conserved, stable, dynamically passive tracer. Broecker and Peng (1982) summarize what is known about most tracers that have been found useful in studying the ocean circulation.

Much of what is taken as conventional wisdom about the long-term time behavior of the ocean circulation is based upon the descriptive properties of chemical tracers. The writings of Wüst are typical of the sort of argument

that is taken for granted by most oceanographers; they are based upon the assertion that the gross distributions of mappable properties permit robust inferences about the overall movement of water in the ocean.

If we assume the WOCE operates a major field program for 3 to 5 years, than with the exception of the chemical tracers, all other observations available can at best describe the ocean during that 3 to 5 year period. Current meter or float or wind observations will yield instantaneous flow or forcing values and annual and interannual means. But we are also interested in making deductions about the ocean circulatuion and its behavior over decades to centuries and longer. Chemical tracers, if we have the wit to use them properly, represent the time integral, that is time average, behavior of the ocean over very long periods of time. Much current activity is directed at making sense of what is often a very complex weighted, in space and time, average. Nonetheless, despite the difficulties, the intense efforts devoted to using tracers in models ranging from simple box balances to elaborate general circulation calculations suggest that the primary limitation at the present time is the paucity of adequate data over much of the ocean.

A useful tracer has several characteristics. It should be easy to measure with adequate accuracy so that it can be mapped in three dimensions (and in time if it is not in steady-state); its chemistry should be simple so that uncertainties in its involvement with global biogeochemical cycles are not overwhelming (tracers such as tritium and fluorocarbons are nearly perfect from this point of view), its sources and sinks should be sufficiently different from other useful tracers that it provides independent information.

Probably no single tracer is absolutely ideal; it is the possibility of observing a large number of them that is the major hope for their use in tightly constraining the large scale time average ocean circulation. A major

component of WOCE is to obtain observations on a global scale of the distribution of enough tracers to infer through appropriate models the long term behavior of the ocean circulation.

7.6 Global Hydrography

Ocean basin scale surveys such as those carried out by the Meteor in 1925-26, and in the North and South Atlantic during the IGY have been the heart of the data base for discussions of the ocean circulation. Such sections serve two purposes--they provide the basic information about water mass distributions and properties, while at the same time giving (through the dynamic method) indications of the flux rates of water masses and, indirectly, their conversion rates.

At the present time, only the Atlantic has been covered by such deep, coast-to-coast surveys and dissatisfaction with the coverage even in the North Atlantic has led to a recent program of augmentation there.

WOCE would include an IGY scale survey of the global ocean, including those regions (South Pacific, Indian Ocean) historically poor in data. Such a survey (which would involve a total of about 10 research ship years spread over the duration of WOCE) simultaneously serves several purposes. 1) It provides the basic climatological distribution of water masses and dynamic topography for the use in circulation models (of very disparate types). 2) In the presence of an altimeter and scatterometer satellites, it provides for the first time a determination of the absolute water flux rates at one moment, 3) Temperature, salinity, and the other chemical properties observed on the lines are the basic observables for making deductions about the nature of the circulation on a time scales greatly exceeding the WOCE lifetime, 4) We provide a baseline for the study, over future decades, for climatological fluctuations in the ocean. 5) Repeated lines provide the essential

information concerning the temporal fluctuations in deep ocean baroclinicity.

Of special importance to WOCE are θ-S distributions where isopycnal surfaces intersect regions of water mass formation by deep convection or wintertime cooling and mixing or where water masses are added by overflows in polar regions. Particular problems exist in defiing an "average" θ-S distribution in regions of sporadic wintertime vertical exchanges. This is of particular concern in the definition of large-scale mixing processes in terms of source water types.

The accuracy and detail to which the interior θ-S fields need to be known depends on whether the primary purpose is model verification or whether it is the estimation of derived quantities such as potential vorticity. An evaluation of the existing θ-S data base for meeting the basic objectives of WOCE is needed.

It is envisioned that the hydrographic program could be conducted on several levels. The research vessels of the primary oceanographic research institutions would provide a standard baseline of the highest possible levels of data quality. In order to cover the world ocean, it is expected that somewhat lower quality data, but more easily and cheaply obtained, would be provided by "secondary" research vessels. Intercalibration between both the primary and secondary data types would have to be very carefully considered.

8. Elaboration on Modelling Foci

Effective treatment of ocean atmosphere interactions in discussions of climate system sensitivity requires quantitative models which are demonstrably realistic enough to be credible for the purposes at hand. Quantitative models of the ocean circulation fall into three broad classes, each with a crucial role in WOCE.

(i) Simple and highly idealized conceptual models with incomplete physics, which have nonetheless captured the essence of some aspect of ocean dynamics or thermodynamics, have made possible major advances in understanding the circulation of the ocean. The complexity of the climate system demands that this approach be pursued on a broad basis, particularly to guide thinking towards critical tests or developments of more elaborate models.

(ii) General circulation models, on the other hand, simulate as many as possible of the relevant competing processes and form the pinnacle of comprehensive descriptions of the system. Because of practical limitations, there are at present two kinds of models: ocean general circulation models (without eddies), and eddy-resolving general circulation models. The first kind are designed to study the global structure of the circulation and the three-dimensional density structure. They include processes of water mass conversion and deep water production; their major deficiency is that they must parameterize crudely a range of physical processes (mesoscale eddies, high-latitude convection, vertical mixing) which may be vital to the climate application. The second kind of model explicitly include mesoscale eddies (through very high horizontal resolution) and has had considerable success in accounting for the geographic structure of eddy variability and lateral mixing. Thus far, however, active water mass formation processes and thermodynamic interaction with the atmosphere have not been included. A marriage of these two kinds of models represents one of the major tasks of the next five years. This will, however, require computing capabilities substantially enhanced beyond those currently

being used. The foci of the observational programme are designed to provide key information necessary for the development and verification of such a complete general circulation model.

(iii) Data assimilation models comprise the third class. With the latest generation of observing systems it is no longer appropriate to analyze data on each observed variable in isolation. Instead, assimilation procedures are required which incorporate, to the extent possible, known kinematic or dynamic constraints between variables. In this process, due regard must be given to the uncertainties and statistical interrelationships of all the observations to produce an integrated picture of the state of the ocean, including inferences of variables which are not directly measurable (such as large-scale vertical velocity). There are a variety of such techniques available at present, known as objective analysis, diagnostic models, and inverse methods. During the coming decade they will become increasingly sophisticated, involving more dynamics and will be the interface in WOCE between the observational systems, on the one hand, and the conceptual and general circulation models, on the other. They provide a vehicle for rational evaluation of data sampling and accuracy, and of the utility of their products as inputs to the models and in model testing and validation.

Carefully fostered periodic intercomparisons between different assimilation models would do much to provide timely feedbacks on data quality control and observing system design issues, and to provide standardized products to focus attention on similarities and differences in model performances.

In addition to the major streams of development discussed above, particular and more specialized modelling needs have been identified for WOCE.

They include (i) development of a highly simplified atmospheric model (or models) suitable for coupling economically to a variety of global ocean models, (ii) assimilation models for geochemical tracers, (iii) isopycnal outcrop models, (iv) other process or regional models with high resolution tailored to represent in detail the physics of particular regions and, (v) mixed-layer models capable of relating to observable signatures of processes such as late-winter convective overturning. In addition, the role of sea ice on the circulation near the ice boundary needs further elucidation.

As general circulation models become increasingly realistic, they will be used in observing systems simulation experiments. For example, whereas critical decisions on the design of satellite-borne scatterometers in orbit in the late 1980s may have to be taken in the absence of simulation experiments, studies of the sensitivity of global models to inaccuracies in the supposed surface wind stress could help in selecting regions for extensive intercomparisons with data from instruments <u>in situ</u> and in developing strategies for coping with directional ambiguities or suspected bias in subsets of the data. Key considerations in such experiments are their relevance to real issues in systems design and the timeliness and credibility of the results. Early involvement by all appropriate parties is essential for a successful outcome.

Developments are also needed in atmospheric assimilation modelling procedures, for the purpose of permitting inferences of surface heat and moisture fluxes over the oceans in places where reliable direct observations are inadequate or unavailable. All relevant information (such as the location of cumulonimbus convection in the tropics, or the field of vertically integrated precipitable water) must be brought to bear.

9. Commentary Upon Some of the Goals

9.1 Oceanic ventilation time and rates of water mass conversion

Ventilation time and water mass conversion are, in part, complementary concepts, which become identical only where advection into the interior of water modified by convection processes is clearly dominant over tracer mixing effects. Moreover, if the net water mass conversion is viewed with respect to a cross section such as the much studied one at 24°N in the Atlantic, it is clear that the "interior" domain may contain subregions of significant water recirculation, with associated mass conversions and related interior (self-balancing) heat flux convergences and divergences.

Nevertheless, it appears that these concepts, used with appropriate caution and with careful definition, provide a useful perspective from which to examine processes related to climate. Two approaches have been identified in WOCE.

First, the introduction of certain anthropogenic materials at the ocean surface provides useful tracers of ocean processes. The observed rates of conversion into the ocean interior, usually primarily along isopycnal surfaces, provide direct estimates of the ventilation in the interior. These estimates depend on adequate information on the evolution of tracer distribution in the ocean interior and tracer concentrations in several regions. At present, these vary greatly in various regions of the World Ocean. Detailed analyses are needed to ascertain what future measurements of tracers will be most useful to meet the basic objective of WOCE. Somewhat similar considerations concerning θ and S are discussed above and the general usefulness of tracers is discussed in Section 4.1 of the Tokyo (1982) Report.

Second, the question of water mass conversion may also be addressed using the indirect physical approach. The best example of this approach is the use

of hydrographic data from the 24°N section in the north Atlantic combined with estimates of the Florida Current transport in estimating the heat flux across that section. This use of section data combined with information on surface topography and direct current measurements in selected regions provides a powerful tool for estimating the transport of various water masses across zonal sections and the estimation of water mass conversion between sections. Detailed examinations of the best location of such hydrographic sections, the proper mix of other direct measurements, the accuracy of the required data sets and other related questions need to be further carried out in the context of the general objectives of WOCE, especially for those regions not now being extensively examined.

9.2 Large-scale signatures of water mass conversion

Water mass temperature-salinity properties are changed by absorption of solar energy and by molecular diffusion. Significant solar heating is limited roughly to the top 100 meters of the ocean, though there is considerable variation with latitude, season, cloud cover, and turbidity of the water. Molecular diffusion occurs everywhere, but has a neglibible effect unless accelerated by small-scale shears associated with three-dimensional turbulence, which locally increases the temperature and salinity gradients. Most models of the general circulation include the assumption that three-dimensional turbulence is uniformly distributed below an energetically turbulent surface mixed layer which varies seasonally. On the other hand, models and observations of the turbulent kinetic energy distribution (which are admittedly limited both in number and global applicability) indicate that the distribution of turbulence is inhomogeneous and intermittent, with the majority of the ocean being in laminar flow at any instant. It is not yet clear how sensitive the general circulation models designed to predict water

mass properties are to this inhomogeneity and intermittency in turbulence. Models designed to predict sea surface temperature and those concerned with the vertical distribution of temperature and salinity in the oceanic thermocline and along the Equator, however, are sensitive to the representation of variations of turbulence.

The design of WOCE should include the collection of information from which to derive key indicators defining large-scale inhomogeneities that affect water mass properties. The choice of such indicators will depend on the detailed models of the processes involved, including solar heating, turbulent mixing in the upper boundary layer, and turbulent mixing in the interior. But because the aim will be to make a global climatological description, there is no question of surveying the processes in detail.

Models of the water mass conversion processes give some indication of how one might proceed to identify key indicators suitable for surveying during WOCE. For example, it may be possible to calculate the profile of solar heating rate from satellite observations of clould cover (surface insolation) and ocean color index (diffuse attenuation index). The key indicators for the turbulent mixing in the upper boundary layer at high latitude may be the net annual Ekman pumping and the annual maximum depth of the mixed layer (in late winter) and the mixed-layer temperature and salinity at that time.

9.3 Seasonal Cycles

The seasonal cycle of heat and stratification is the largest signal in the upper ocean and has considerable inter-annual variability. Its importance to WOCE lies in the fact that water mass conversions are highly seasonally dependent. Thermocline ventilation is controlled by the maximum winter depths of the convective mixed layer. In polar regions, convection and formation of

sea ice may be strongly influenced by variations in the salinity of the mixed layer. The extent of deep winter convection may be affected by pre-conditioning of the surface layers by mesoscale eddies.

Within the context of WOCE, some regions with known strong annual air-sea interactions will require seasonal observations of the upper ocean structure with instruments *in situ*. Globally, such regions can only be inferred from satellite observations through model response studies. That, in turn, requires a significant effort to improve upper-ocean modelling for climate studies.

The seasonal variation of gyre-scale surface geostrophic circulation has never been adequately observed, and observations which constrain models of this with an accuracy of ±1 cm/s (±10 cm sea level/1000 km) would provide important new information.

9.4 Variability

Little is known about the climatological spectrum of variability of the ocean over the WCRP time band (several weeks to several decades), and virtually nothing is known about the regional variation of the spectrum. It is assumed that the spectrum is broadband, with spectral peaks at the annual cycle and its harmonics, and that significant energy will still be present even at longer periods. There will, therefore, be significant variability on timescales both shorter than and longer than the duration of WOCE.

The longer-term variations will constitute a non-stationarity of the system sampled by WOCE as a "snapshot" (albeit with a shutter open for 5 years). There is a need to establish a long-term monitoring programme that will indicate how the WOCE snapshot relates to longer-term variability. Such monitoring will have to extend over a period an order of magnitude longer thain WOCE, and concentrate on a few key variables in a few key locations.

Careful study must be given to the specification of such monitoring, and determining its details must be a goal for WOCE.

Variability on timescales shorter than the duration of WOCE can, in part, be resolved, but the remainder will contaminate the WOCE data set by aliasing or by non-synoptic distortion. Some observing systems, such as those based on satellites, will offer many global surveys of the required measurements during the period of WOCE; the variability will be resolved within the spectral window of such observations. Other observing systems, such as hydrographic sections that are collected only once in WOCE, alias or distort the variability on periods shorter than their own duration, and represent a (distorted) snapshot that samples the variability at one time within the duration of WOCE. Exploratory time series measurements will be needed to determine the degree of sampling non-representativeness of such observations, and the degree of error in them due to aliasing.

Sea Surface Topography--Toward A WOCE Strategy

Carl Wunsch

14 June 1983

1. Introduction

Much of the stimulus for discussing the possibility of conducting a global circulation experiment comes from the potential availability of an altimetric satellite of high precision and accuracy. Unlike any other observable, surface topography is an oceanic variable, measureable from space, that is dynamically related to the three dimensional ocean circulation at great depths. The surface elevation represents a direct boundary condition on the quasigeostrophic interior flow fields of the ocean. This is not so for surface temperature, roughness, dielectric constant..., all of which are determinable from space, but whose relationship to the flow field at depth is very obscure and in some cases extremely complex.

The uses of a high accuracy altimetric satellite were discussed at some length by The Topex Science Working Group (1981) and that discussion will not be repeated. Instead, we simply point out that Seasat lasted just long enough to demonstrate quantitatively what could be done (it did not last long enough to teach us anything really new about the ocean). The Seasat mission has thus given rise to a considerable literature which we will summarize below.

But the direct measurement of sea surface topography and its use for making inferences about the ocean circulation has a history long ante-dating spacecraft. Measurements of surface elevation by tide gauges are the only really long direct measurements we have of the ocean. In the hands of a number of investigators, these observations have been extremely useful for studying large scale, low frequency behavior. It behooves us to be careful that this very effective observational tool not disappear. In fact, the

strategy proposed here for altimetry, is based upon the supposition that the existing global tide gauge network will remain in place and, indeed would be supplemented, in order to maximize the effectiveness of the altimetric measurement.

2. What Good is Altimetry?

A. Oceanic Variability

Many investigtions with Seasat (and the prior Geos-3) mission have already been published in the open literature. (Fu 1983a has written a review, but it is as yet not published). To give some flavor of what has been learned consider figure 1 from Cheney, Marsh and Beckley (1983). It shows an estimate of global mesoscale variability as determined by Seasat from one month of data. This paper (and others, e.g. Douglas and Cheney, 1981, Menard, 1983) compare the results to other such estimates (e.g. Wyrtki et al., 1976) from conventional means and discuss the reasons why the altimetric measurement is probably more representative (keeping in mind the extremely short Seasat data base).

Figure 2 is taken from Fu (1983b); he showed that estimates could be made not only of the total mesoscale variability energy but also the frequency/wavenumber characteristics of that variability as a function of position.

Seasat demonstrated that to the extent global statistics of the mesoscale variability remain of scientific interest, that an appropriate future altimetric satellite could essentially solve the problem (I am aware in this statement, and others like it below, that there are no sweeping generalities that are actually valid--one must consider separately for example, variability in the form of deep topographically trapped waves with no surface geostrophic pressure signature.) Many have argued, that for this reason alone, flying such a

SEASAT ALTIMETER MESOSCALE VARIABILITY

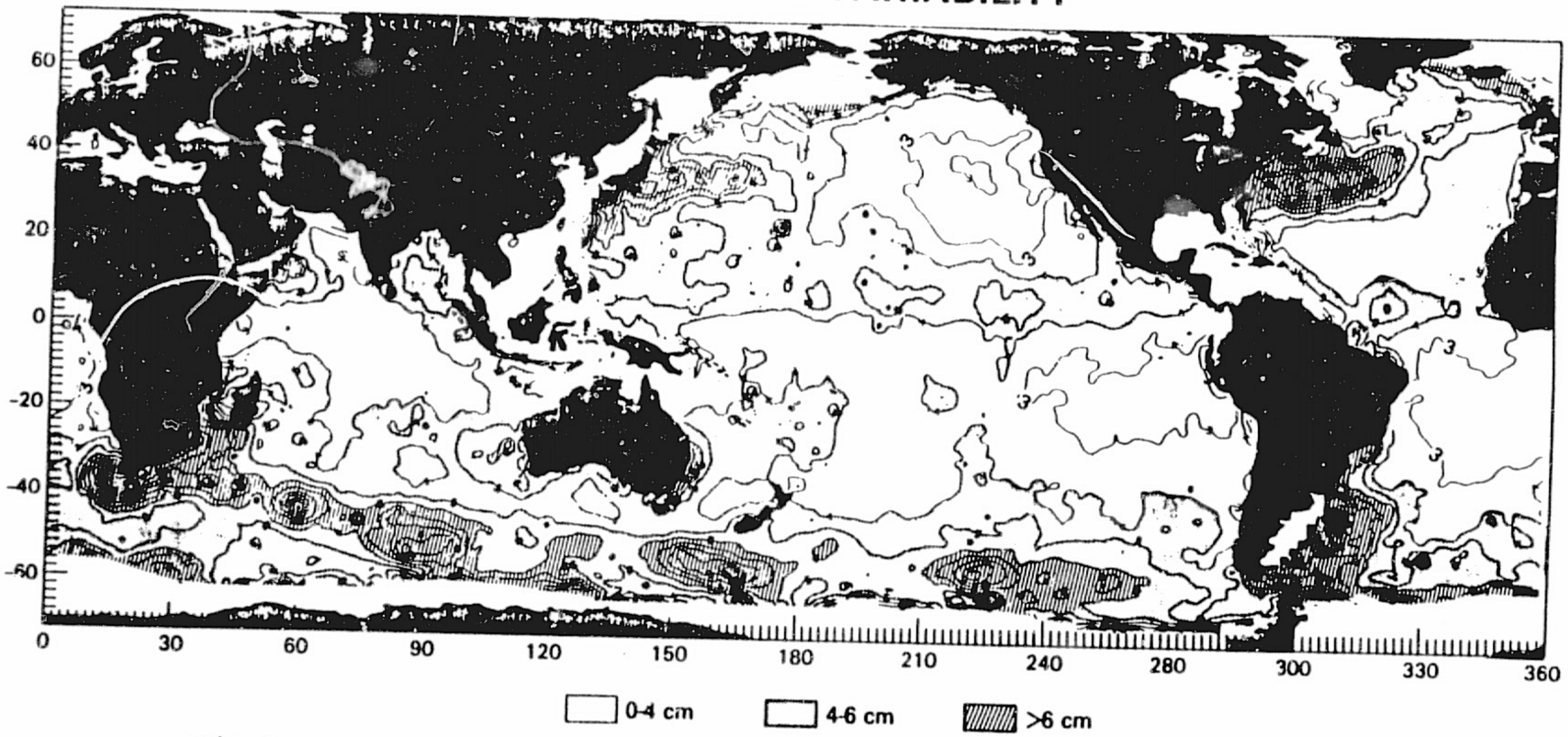
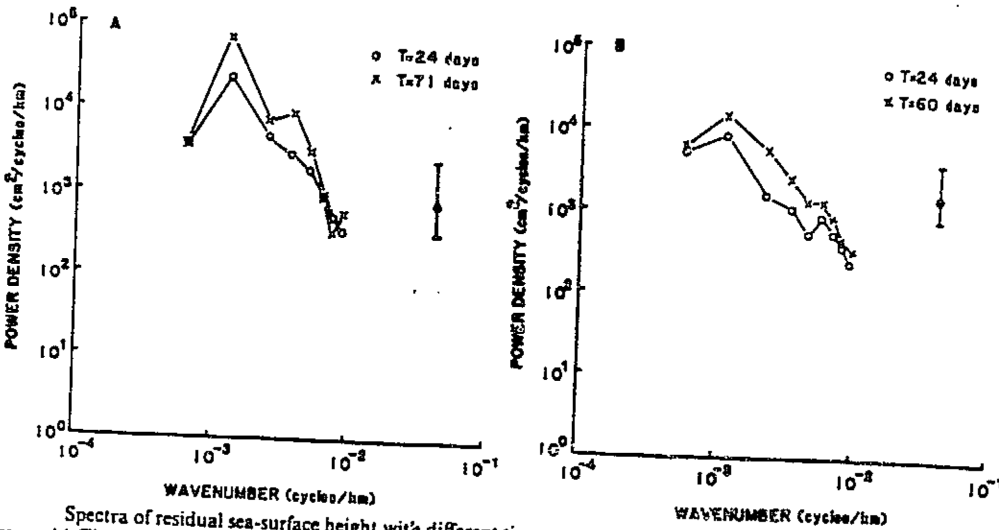
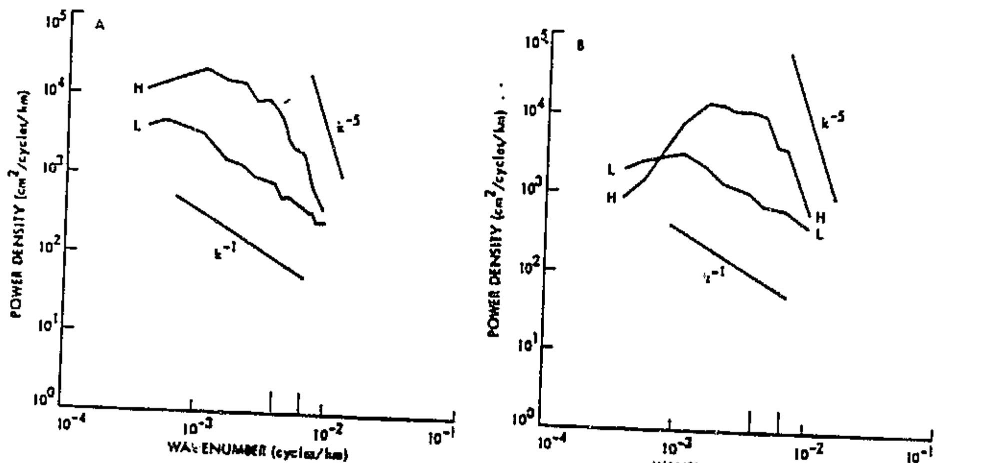


Fig. 1 Global mesoscale sea height variability measured by the SEASAT altimeter, September 15 to October 10, 1978. This map was constructed from 110,000 globally distributed variability values determined every 7 km along the tracks shown in Figure 1. A gridding routine was then used to obtain smoothed values at regular 2° intervals of latitude and longitude (see Appendix A). These were contoured to reveal large-scale variability patterns due to current systems. The North Atlantic and North Pacific are dominated by the highly energetic Gulf Stream and Kuroshio systems that extend seaward nearly 4000 km. In the southern hemisphere the Agulhas Current below Africa and the Falkland/Brazil Current confluence off South America are clearly apparent. High variability due to the Antarctic Circumpolar Current extends in a nearly continuous band around the polar oceans, with isolated maxima coinciding with major topographic ridges and plateaus. Owing to the predominance of values less than 4 cm in mid-ocean, the north equatorial current systems in both the Atlantic and Pacific can be seen as zonal bands of higher variability.

(Cheney et al, 1983)



Spectra of residual sea-surface height with different time spans (denoted by T) along the track of (*a*) group 62 and (*b*) group 14. The error bars represent 95% confidence limits for the spectra with longer time spans, which have larger error bars.



(*a*) The average spectra of residual sea-surface height for the high-energy areas (labeled by H) and the low-energy areas (labeled by L). (*b*) The corresponding scalar-wave-number spectra. The two large tick marks on the wave number axis indicate wavelengths of 150 km and 250 km, respectively.

Fig. 2 (upper). Wavenumber spectra from two different overall durations and areas from Seasat altimetry. More energy is contained in longer period records.

(lower) Wavenumber spectrum from high and low energy regions of ocean from altimetry.

(Fu,1983).

mission would be well worthwhile-circumventing the now painful, and expensive deployment of moorings or floats all over the world.

Of course, the variability transcends the meso scale. At the present time it is almost impossible to answer the question of what is the nature, magnitude and geographical distribution of oceanic variability on spatial scales exceeding order 200 km and time scales exceeding a few months (something is known in the tropical Pacific largely from the tide gauge network there). In that sense, an altimetric satellite would permit for the first time, an ability to answer the question of whether the ocean does vary at all on these scales (which include the extremely important annual cycle).

(As an example of what this would mean, consider the note by Worthington (1977). Based upon a single hydrographic section across the Gulf Stream in 1977, Worthington suggested that after the extremely cold winter of 1976-1977, that the Gulf Stream was carrying more water than normal (Leetmaa has questioned this interpretation of the data, but it is the raising of the question that we are interested in). Suppose Worthington is right; was this increase in transport something that affected the entirety of the gyre? Was it perhaps confined to the recirculation south of New England? What does the normal cycle of transport look like? How did the supposed increase decay with time over the gyre? At the present time we could not begin to answer these questions. But a properly designed altimeter mission would go a long way toward providing answers).

Does the ocean have any interannual variability at all on time scales of a few years? Good maps of the surface elevation could tell us. The climatological consequences of the answer are very important.

B. What Would One Do With Variability Data?

Some of our questions about the variability are like the ones described above--e.g. is there any variability on scales larger than the mesoscale--does the gyre "wobble" as Stommel and Armi have suggested? The impact of a better descriptive oceanography of the variability would be very great and should not be minimized. But it is one of the major virtues of altimetry that we also can see how to quantitatively use the data.

Figure 3 is one of Holland's (private communication) EGCM runs. It is displayed here as a representative of all such models now being developed and their anticipated offspring 5 to 10 years hence. As Schmitz and Holland (1982) and Bretherton (private communication) have emphasized, the ability of EGCM's to reproduce the basis eddy statistics (energy levels, frequency and frequency/wavenumber spectra) is a very stringent dynamical test of the models. To the extent that such models get the eddy statistics correct, we tend to have a greater faith in their ability to compute other, perhaps more interesting, but unobserved phenomena (like the oceanic heat flux).

A more direct dynamical approach was outlined by Wunsch and Gaposchkin (1980) and Munk and Wunsch (1982, Appendix D, reproduced here as Appendix A). It is shown that in principle one can go from an altimetric varibility measurement directly to inferences about the isopycnal variability and hence to statements about the three dimensional variability. Obviously the EGCM's can do this in a more sophisticated way than one can do with a simple analytical model used in the appendix, but the principal is the same.

How accurately can all this be done? The question has been discussed by the Topex Science Working Group, 1981, Fu, 1983b and by Wunsch and Zlotnicki (1983, unpublished, attached here as Appendix B). When one deals with any two dimensional time varying field, the answer is not reducible to a single

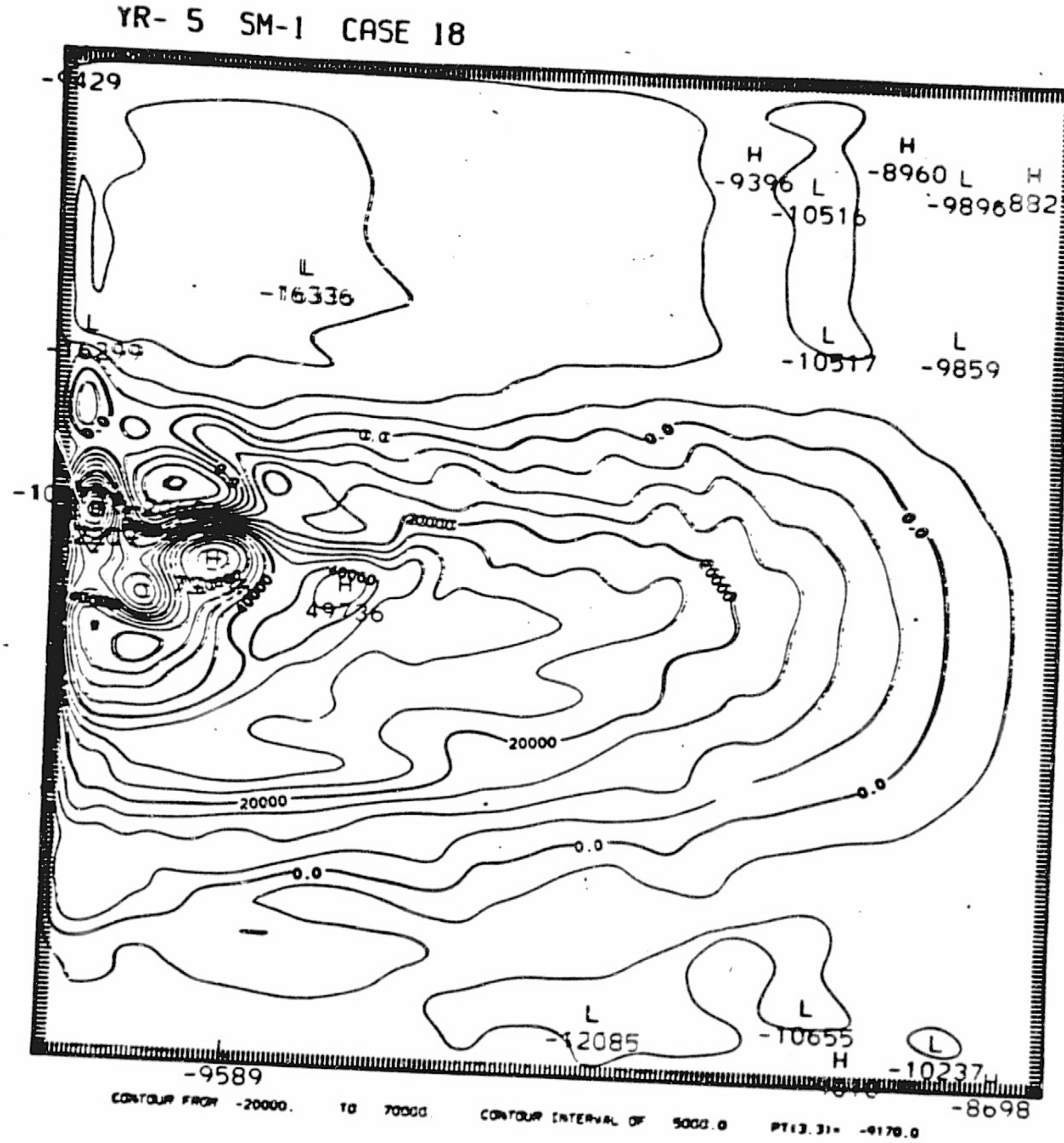


Figure 3. One year average of upper layer of an EGCM(Holland).

number, but is a function of frequency and wavenumber, averaging time, system accuracy, position, and calibration procedures. Taken altogether, and with a sensible calibration procedure (outlined later), we believe that all spatial scales from the width of the Pacific Ocean (10,000 km) down to the baroclinic Rossby radius of deformation (about 30 km) will be observable at the sub-centimeter level on time scales from order 20 days to 5 years.

C. The Time Average Circulation

Here again, the results from Seasat are probably more convincing than any argument. Figure 4 is taken from Tai and Wunsch, 1983 (the paper is attached as Appendix C). It shows a determination of the absolute dynamic topography of the ocean as a three month average over the lifetime of Seasat. This result differs from more familiar pictures in that a spatial filter has been applied to remove all wavelengths from the surface that are uncertain because of uncertainties in the geoid. For comparison, Figure 5 is the recent Levitus' (1982) global picture from averaged hydrography run through the same low-pass filter. The paper, Appendix C, discusses how these surfaces are generated, and why the Levitus and satellite pictures should be different in some respects.

At the present time, the construct ion of charts such as that shown in Figure 4 is confined to either the very largest scales as shown, or to comparatively small areas where the entirety of the wave number content can be mapped. The reason is lack of certainty about the underlying gravitational equipotential surface.

Considerable thought must be applied to developing a strategy for working with such filtered or spatially limited data. We propose a two-fold approach. First, the manuscript attached as Appendix D discusses the use of time averaged altimetric surfaces with limited geoids in the context of ocean

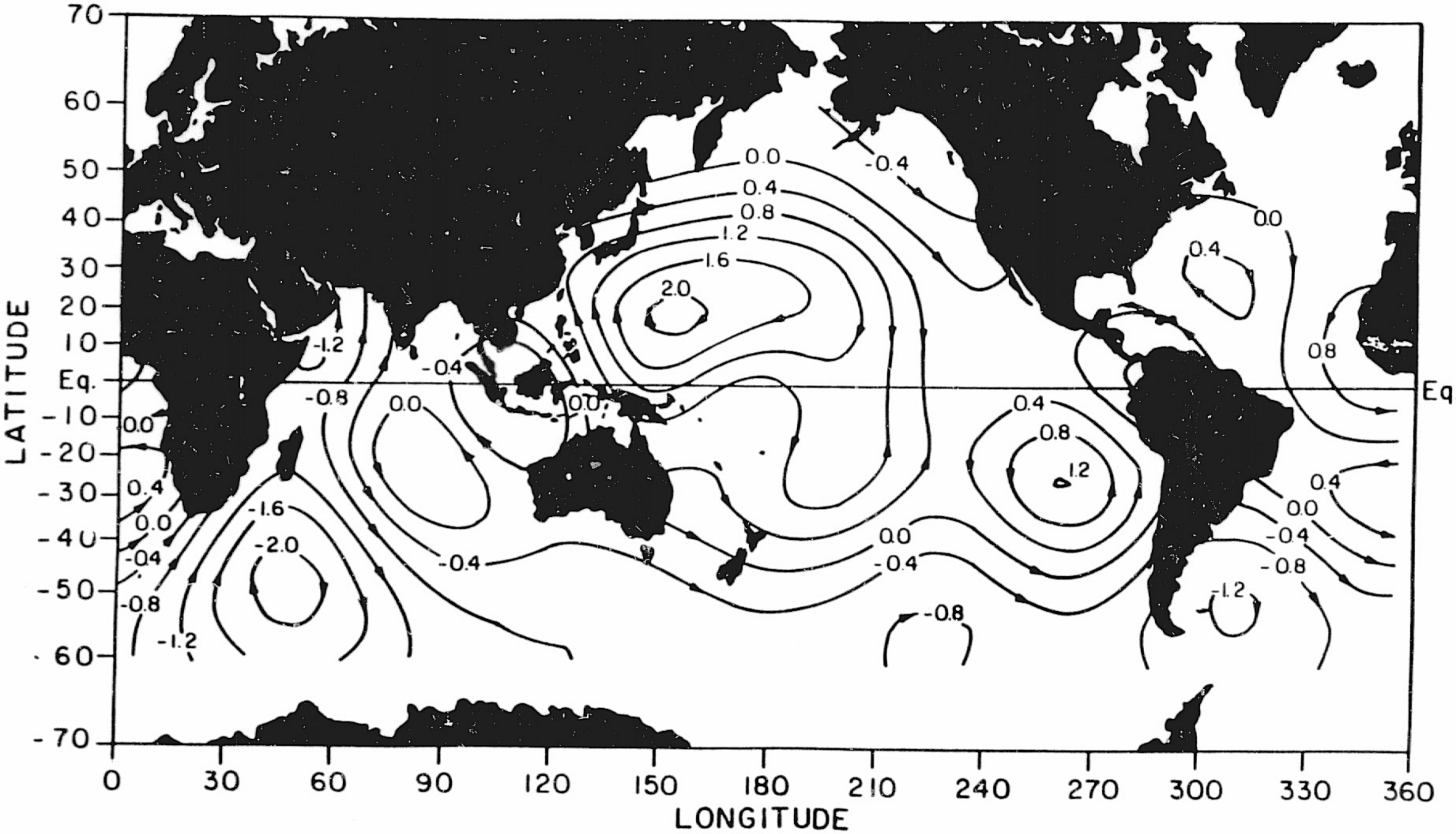


Fig. 4. Absolute dynamic topography of the world ocean, wavenumbers below spherical harmonics of degree and order 6, from 3 months of Seasat altimetry (Tai and Wunsch,1983).

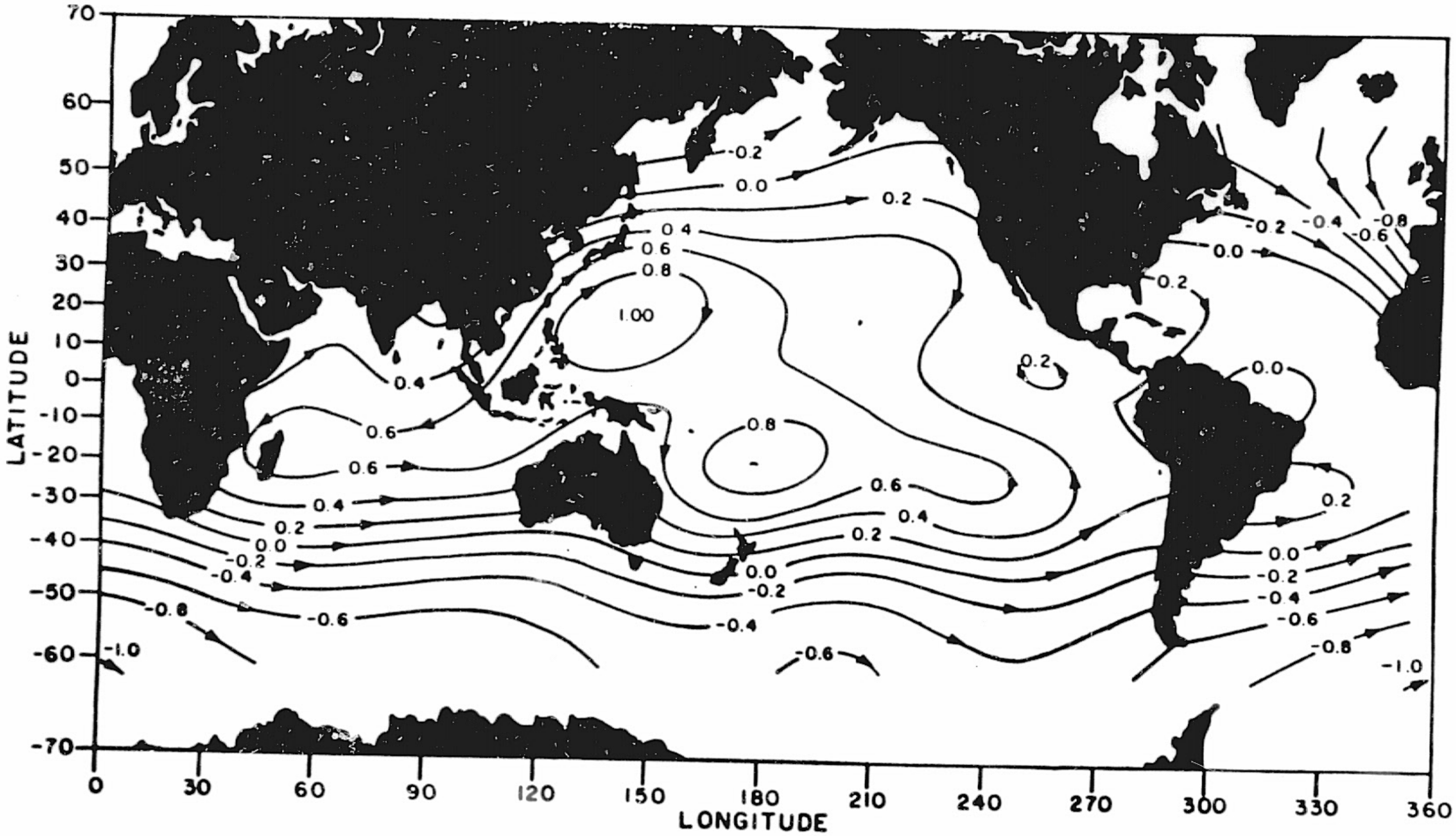


Fig. 5. Average dynamic topography of world ocean, obtained from historical shipboard data (Levitus,1982), and filtered in same way as figure 5.

circulation models. The fundamental notion is that known dynamics permit one to systematically constrain the ocean circulation from filtered or geographically limited measurements--given the property of fluids that all spatial scales and all geographical regions tend to be linked dynamically and kinematically.

The second component of the strategy is to work for ultimate improvement in the geoid. Special purpose spacecraft have been designed (e.g. GRM, nee Gravsat) which should move the useable shortwavelength cutoff down to about 200 km (Breakwell, 1979) thus leaving indeterminate only those scales between 200 km and the Rossby radius. If we can identify critical oceanic regions (e.g. the western boundary currents) where determination of the absolute geostrophic flow would produce especially important constraints, we can consider obtaining adequate data to produce regional geoids of high accuracy (e.g. Marsh and Chang, 1978). Notice furthermore that in situ determination of the geostrophic flow at one instant during the flight of an altimeter satellite, locally determines once and for all the local geoid slope (which is the dynamically important variable). It is possible that a strategy for geoid improvement by in situ observations of flow over comparatively short periods of time would be highly effective.

D. What Should One Do With the Mean Altimetric Data?

The manuscript attached as Appendix D sketches the quantitative use of mean altimetric measurements in conjunction with models and with a great variety of other observations including especially wind stress measurements. In general such observations provide strong constraints upon circulation models (although more elaborate simulations need to be performed). Figure 6 is an example of the operation of such constraints. It was produced in the following way. The zonal hydrography of the North Atlantic from the IGY was

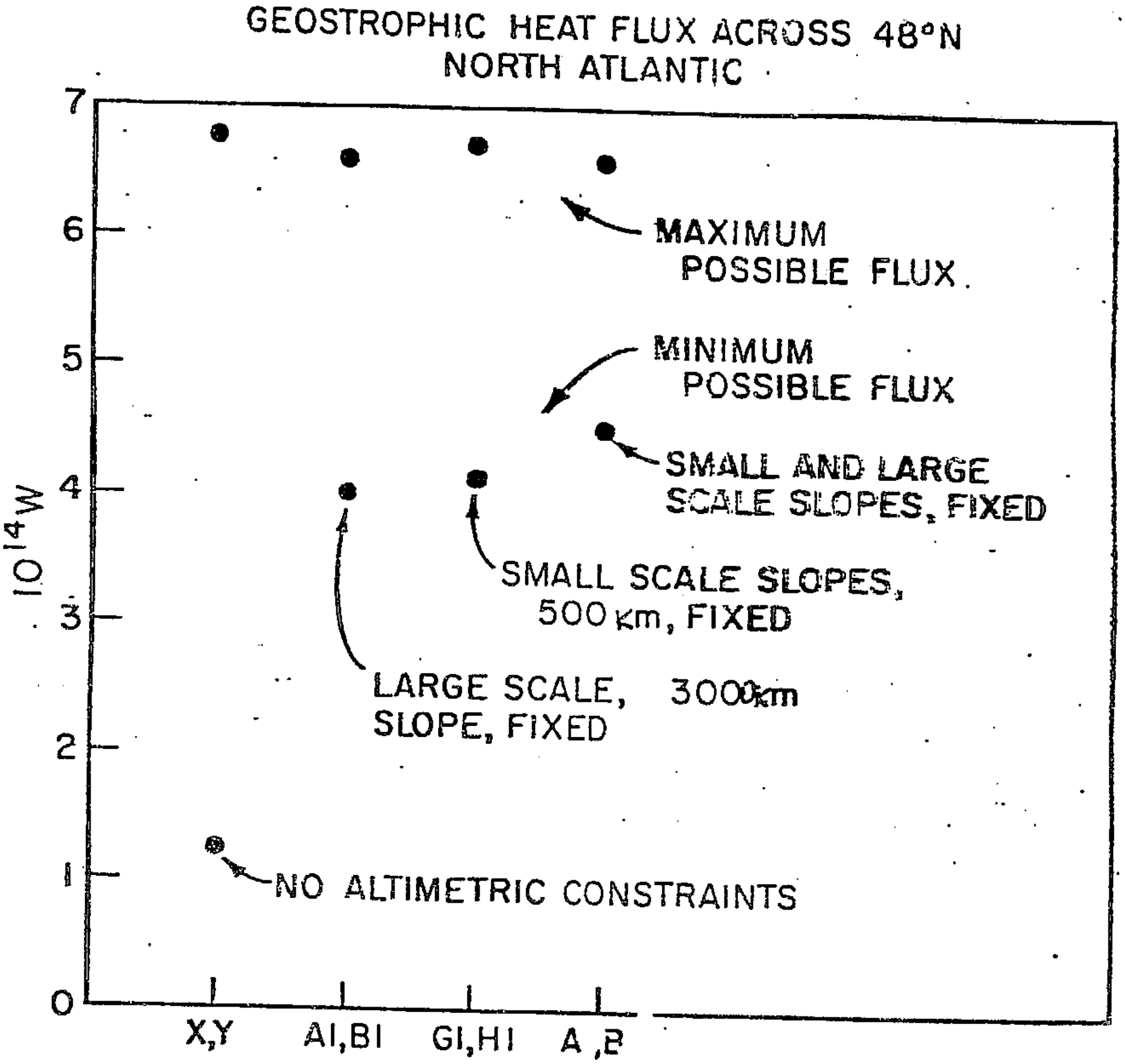


Figure 6. X,Y represent the range (realistic) of meridional heat flux across 48N in the Atlantic using hydrographic data only. A1,B1 show the reduction in range when the slope over 3000km is imposed (this is a simulation). G1,H1 show what the possible range is when the slope is fixed over 500km. and finally A ,B show what happens when both slope scales are fixed.

used to write a series of constraints (Wunsch, 1978) useful for determining the unknown reference level in the thermal wind. Using a linear programming scheme (Wunsch, 1983b) the maximum and minimum values of the poleward flux of heat across 48°N were sought (the question--what is the range of heat flux across this latitude?-was chosen as an example of a climatologically interesting problem). The range for this purely hydrographic data set is displayed on the left of figure 6. It was then supposed that an altimetric measurement was available, that provided the slope of the sea surface across the Atlantic at 48°N only and only on the 3000 km scale (an arbitrary slope value was chosen). A new set of bounds was then computed, shown next in the figure. It was then supposed that only the slopes on a 300 km scale were available, and finally that both the 300 km and 3000 km slopes were measured. One sees from the figure the considerable narrowing of the range as plausible altimetric information is added. In the end, the range may be so small that the remaining uncertainty is no longer a problem. Much more exploration and understanding of the constraining value of sea surface slopes with more sophisticated models needs to take place, but I believe the principle is clear.

3. An Overall Strategy

Figure 7 shows some of the tide gauge network displayed by Wyrtki (1979) roughly superimposed upon the expected 10 day coverage by Topex. I have added both existing and potential positions for tide gauges in other oceans. This network would provide the major form of calibration for Topex or any other altimetric satellite. The calibration issue is an important one. Although there are many sources of error in altimetric measurements (see Wunsch and Gaposchkin, 1980, Topex SWG, 1981, Fu, 1983b) most of them (water vapor, ionospheric electron content, atmospheric load, etc.) are correctible to a

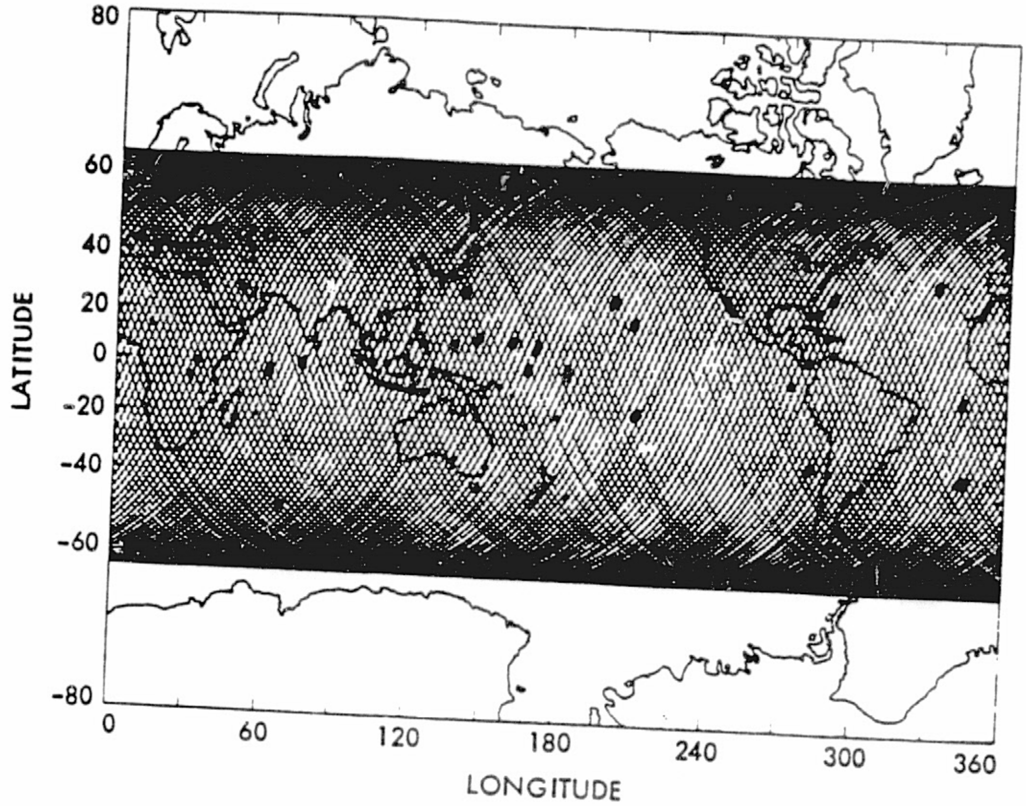


Fig. 7. Some of the existing or hypothetically available locations for island tide gauges superimposed upon the 10 day coverage by TOPEX.

very high accuracy. The most important remaining time dependent error is due to orbital uncertainty and occurs principally on the longest scales (an orbit diameter). The Topex Precision Orbit Determination Group has estimated that this error can be reduced, through realistic tracking systems, to order 14 cm in each ground track (it was about 50 cm in Seasat). Much of this error will be reduced by simple temporal averaging of successive repeating passes to effect a $\sqrt{N}$ reduction. But the most important procedure for ultimate error reduction will almost surely be the so-called crossing arc method (e.g. Rapp, 1983) where the long wavelengths in successive arcs are adjusted to make the altimetric measurements agree where different arcs cross each other. Wunsch and Zlotnicki (1983, in preparation) have studied the remaining residual errors. Figure 8 shows the wavenumber spectrum of the error in an altimetric surface, taken along one of the measurement arcs after an adjustment (effected in this case through an objective mapping procedure). It was assumed, extremely pessimistically, that the tracking error gave an uncertainty in each orbit of 1 meter, not 14 cm, and that the arcs were 1400 km apart, not 200 km as they actually would be). The rms error is reduced from $(100\ \mathrm{cm})^2$ to $(24\ \mathrm{cm})^2$ and is very small at short wavelengths. If the actual value of the sea surface is fixed at one point along a track (by a tide gauge for example), the rms error along any other arc was reduced to about $(12\ \mathrm{cm})^2$. It thus appears that the global tide gauge network will play a crucial role in error reduction over the entire globe (because the resulting altimetric surface is a global one). In this sense, one might prefer to think of the altimeter as an interpolator between surface measurements by tide gauge. In the next several months we need to quantify the global error statistics and decide where to argument the existing island network.

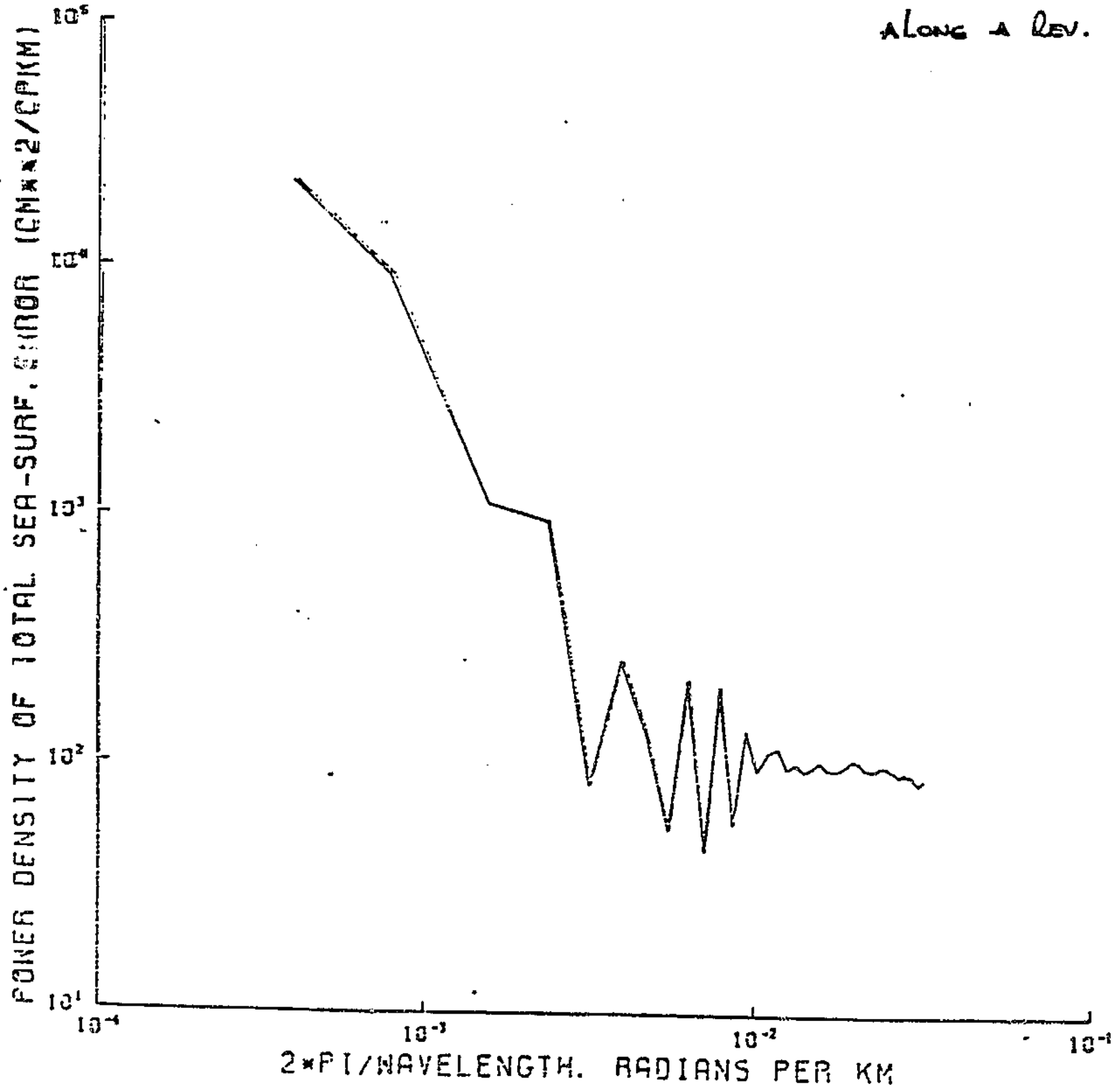


Fig. 8. Wavenumber spectrum of error in an altimetric surface. Area under curve is equivalent to an rms error of $(24cm)^2$. Actual error in practice would be much less than this both because the amount of data used here was a small fraction of that anticipated, and because a very pessimistic (1meter) orbital error was assumed.

It is proposed therefore that WOCE should be built around a suitable altimetric satellite--one with adequate accuracy for improving on what we already know about the ocean, and one that will fly long enough to provide adequate statistics on the variability, to produce a long enough average for studying the interaction between mean and fluctuations, and long enough to mount shipboard measurements, serially, over the world ocean. The WOCE scientific groups should encourage programs for geoid improvement to be ready at the time when a suitable altimetric average becomes available; maintenance of the global tide gauge network; a program of tidal improvement both for understanding the tides and reducing them as a source of altimetric error; and the development of models suitable for "assimilating" altimetric data.

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Appendix A. Inferences about the interior ocean from surface measurements by satellite

Consider a region of the ocean where linearized theory applies well to the time-dependent motions. Then over a flat sea floor the pressure field describing the three-dimensional fluid motion may be written (using F' for dF/dz)

$$p_r(x,y,z,t) = \lambda_r F_r'(z)\, P_r(x,y)\, e^{-i\omega_r t}. \tag{D 1}$$

The vertical wave functions $F_r(z)$ can be obtained from

$$F_r'' + (N(z)/\lambda_r)^2 F_r = 0 \tag{D 2}$$

subject to the boundary conditions

$$F = 0 \quad \text{on} \quad z = -D, \quad F' - g\lambda_r^{-2} F = 0 \quad \text{on} \quad z = 0, \tag{D 3}$$

where $N(z)$ is the buoyancy frequency, thus resulting in a Sturm–Liouville problem for linear Rossby waves. P satisfies a horizontal equation

$$L(P_r(x,y), \lambda_r) = 0, \quad P_r(x,y) = Y_r(y)\, e^{ik_r x}. \tag{D 4}$$

Note that each mode has its own characteristic wavenumber and frequency. If we can deduce ω_r, k_r from an altimetric measurement, we can determine $P_r(x,y)$ from the differential equation (D 4), and similarly F_r from (D 2) and (D 3). The pressure field is given by the mode summation

$$p = \sum_r p_r = \sum_r d_r F_r(z)\, Y_r(y)\, e^{i(k_r x - \omega_r t)} \tag{D 5}$$

and the perturbation density field ρ is found from

$$-\rho g = \partial_z p$$

and is thus completely determined from the altimetry. More complex but similar models can be used in regions of strong mean shear or bottom topography.

The *variable* velocity at depth can be inferred from a mode summation

$$\begin{Bmatrix} u_r \\ v_r \end{Bmatrix} = F_r'(z)\, e^{i(k_r x - \omega_r t)} \begin{Bmatrix} U(y) \\ iV(y) \end{Bmatrix}$$

which is analogous to (D 1) for pressure (density), and subject to the same considerations.

In determining the time average density and velocity field, the practical approach is probably to use a numerical model and assimilation techniques. But an interesting analytical possibility is to relate the measurements to the thermohaline circulation models of the ocean (Welander 1971). According to Welander, there is an integral of the quasi-geostrophic equations of motion in the form

$$\sin(\text{latitude})\, d\rho/dz = G(\rho, p + \rho g z),$$

where G is an arbitrary function. Under some circumstances the satellite determination of surface pressure and stress (and hence Ekman suction) could provide information on the three-dimensional behaviour of G and hence of ρ, and the three components of velocity.

Although this is an interesting possibility o be explored, the existence of the mesoscale and other ocean variability raises fundamental questions. The variability induces eddy correlations not appearing in the models. Measurements described in this paper could finally answer the question of whether the resulting eddy fluxes are important to the time-average flows. There is probably no substitute for good eddy-resolving general circulation models until the eddy processes are better understood.

APPENDIX B (Available Later)

APPENDIX C

An Estimate of Global Absolute Dynamic Topography

Chang-Kou Tai [1]

and

Carl Wunsch

Center for Meteorology and Physical Oceanography
Department of Earth, Atmospheric and Planetary Sciences
Massachusetts Institute of Technology
Cambridge, MA 02139

ABSTRACT

We estimate the absolute dynamic topography of the world ocean from the largest scales to a short wavelength cutoff of about 6700 km for the period July through Sep ember 1978. The data base consisted of the time averaged sea surface topography determined by Seasat and geoid estimates made at the Goddard Space Flight Center. The issues are those of accuracy and resolution. Use of the altimetric surface as a geoid estimate beyond the short wavelength cutoff reduces the spectral leakage in the estimated dynamic topography from erroneous small-scale geoid estimates without contaminating the low wavenumbers. The resulting surface represents a current best estimate of the long wavelength components of the surface boundary conditions of the oceanic general circulation.

[1] Now at Scripps Institution of Oceanography, La Jolla, CA 92093

1. Introduction

Oceanographers have long recognized that knowledge of the absolute shape of the sea surface elevation, ζ, relative to an equipotential surface of the earth's gravity field (the special one usually denoted the "geoid") would have a profound impact upon the problem of determining the global ocean circulation. The slope of the surface elevation is equivalent to the determination of the unknown reference level velocity in the dynamic method. If one knew ζ with adequate accuracy, the notorious level of no motion problem would disappear from the concerns of oceanographers.

The advent of satellite altimetry has for the first time made it plausible to discuss seriously the possibility of direct determination of ζ. The concept has been described by Mather, Rizos and Coleman (1979), Wunsch and Gaposchkin (1980), Roemmich and Wunsch (1982), Tai and Wunsch (1983), Tai (1983) and others. We will not reproduce here all of the many details, but will briefly mention the major conceptual problems. Consider the absolute shape of the sea surface $S(\theta,\lambda)$ as seen from a coordinate system fixed at the center of the earth (θ, λ are latitude and longitude). We may write it as

$$S(\theta,\lambda) = \zeta(\theta,\lambda) + N(\theta,\lambda) + r(\theta,\lambda)$$

where $N(\theta,\lambda)$ is the geoid and r contains a number of contributions (discussed at length by Wunsch and Gaposchkin, 1980 and Fu 1983) which we will consider to be noise for present purposes.

From the measurements of Seasat, highly accurate estimates, $\hat{S}$, of S have been produced (see Marsh and Martin, 1982; Rapp, 1983) averaged over the period July to September 1978. Coupled with an estimate, $\hat{N}$, of N, (where $N = \hat{N}+\delta\hat{N}$) we obtain an estimate $\hat{\zeta}$ of ζ as

$$\hat{\zeta} = \hat{S} - \hat{N} = \zeta + \delta\hat{N} + r \tag{1}$$

(Wunsch and Zlotnicki, 1983, show that $|\delta\hat{S}|<<|\delta\hat{N}|$, in general)

but note that crossing-arc analysis (Marsh and Martin, 1982) has been used to reduce the orbit uncertainty error in constructing $\hat{S}$. But systematic errors cannot be removed this way. Suppose the gravity field estimate which was used to determine the orbit underestimates the gravity field in a certain region. The true orbit is then lower over this region than the computed one. The result is that estimated sea level $\hat{S}$ is higher than true sea level over this region. Further compounding of the problem occurs if one uses the same geoid estimate in eq (1) to take the difference. We should also point out that Lambeck and Coleman, 1983, have challenged published statements of geoid accuracy without resolving the issue.)

To make the expression (1) of any use, one must estimate N accurately. At the present time unfortunately, $|\delta \hat{N}| > |\zeta|$. Zlotnicki (1983) reviews the various procedures (note that for many geophysical purposes, where N is the quantity of direct interest, it often suffices to write, to the lowest order $\hat{N} = \hat{S}$, or $\hat{\zeta} = 0$. This approximation is inadequate for studying oceanic flow fields).

As with any function defined on a sphere, we may write

$$N = \sum_{n=0}^{\infty} \sum_{m=-n}^{n} N_n^m \, Y_n^m(\theta, \lambda) \tag{2}$$

where the N_n^m are spherical harmonic coefficients. For low degree and order (small n,m) the coefficients N_n^m have been determined by tracking earth orbiting satellites over the past 25 years. Because each term in (2) (more precisely the related term describing the equipotentials at satellite height rather than at sea level) perturbs a satellite orbit somewhat differently, one can obtain estimates $\hat{N}$ for small n,m from such tracking data (see Kaula, 1966).

Because the gravity field above the surface of the earth is a solution to Laplace's equation, the effects on satellite orbits of terms for large n,m diminish rapidly, and above degree n of about 20, the effects are so insignificant as to render impossible the reliable estimation of N_n^m.

These high degree and order terms must be estimated locally from shipboard and continental measurements of gravity acceleration, or by indirect means (which we will discuss elsewhere). We have global estimates of $\hat{N}$ for low degree and order, but have only regional estimates for high degree and order.

Spherical harmonics of low degree and order correspond to the small

wavenumber components of N (the wavelengths are approximately 40,000 km/n). Preliminary estimates of the accuracy with which N is known now suggested to us that although we could not usefully employ (1), that a modified form might be extremely powerful.

Let $h' = f**h$ represent the two-dimensional filtering operation, with filter f, designed to remove from any field h all wavenumbers above a cutoff $k = k_c$. On a sphere, the filter is designed to remove all spherical harmonics from the field h above degree $n = n_c$ (the filter used is a straightforward spherical harmonic expansion with the expansion cut off at degree n_c). Then we can attempt an estimate

$$\hat{\zeta}' = (\hat{S}-\hat{N})' \tag{3}$$

of the spatially low-passed ζ. At the end, we will discuss the utility of such estimates.

2. Procedures

Lerch et al. (1979) produced an estimate of the geoid from orbital perturbations which they denoted GEM-9 (for Goddard Earth Model -9). We write their estimate as $\hat{N}_1 = N-\delta\hat{N}_1$ where $\delta\hat{N}_1$ is the error. Table 1 lists their estimate of the rms error $\delta\hat{N}_1$ as a function of degree (the square root of the power density spectrum of $\delta\hat{N}_1$). Tai (1983) analyzed the energy in Wyrtki's (1975) estimate of the dynamic topography of the Pacific Ocean, an estimate which was based solely upon hydrographic data and a 1000 decibar reference level. He found that of the power in Wyrtki's surface lying between degrees 1 and 36 (wavelengths of about 1100 km), over 80% was contained in degrees 1 to 6 (about 6700 km), thus reinforcing the idea that the oceanographic signal might be visible at long wavelengths in the difference (3).

Tai and Wunsch (1983) displayed the surface (3) for the Pacific Ocean complete to degree and order 6 with encouraging results. We have thus been led to make a similar attempt for the entire globe which is the purpose of this note.

The reader may have noticed that although the geoid is defined globally, the sea surface elevation is not and that altimeter measurements are meaningful only over the oceans. The complications this brings into the problem are discussed by Tai (1983); in summary, the problem is treated as follows. We can define ζ to be anything we please over land and it is most convenient to define it as zero there. The spherical harmonic expansion of ζ is thus defined as that of a function equal to the sea surface elevation over water, and zero everywhere else. Let ζ_o be the ocean function as defined by Munk and MacDonald (1960), i.e. we have

$$\begin{aligned}\zeta_o(\theta) &= 1 \text{ over ocean}\\ &= 0 \text{ over land}\end{aligned}$$

Eq. (1) multiplied by ζ_o becomes

$$\hat{\zeta} = \zeta + \delta N\zeta_o + r\zeta_o.$$

Let $\overline{\delta N} = \delta N\zeta_o$ and $\overline{r} = r\zeta_o$. If we expand $\hat{\zeta}$ in spherical harmonics,

$$\hat{Z}^m_n = Z^m_n + \overline{\delta N}^m_n + \overline{R}^m_n,$$

where $\hat{Z}^m_n$, Z^m_n, $\overline{\delta N}^m_n$ and $\overline{R}^m_n$ are the coefficients of $\hat{\zeta}$, ζ, $\overline{\delta N}$, and $\overline{r}$ respectively. The coefficients $\overline{\delta N}^m_n \neq \delta N^m_n$; rather they are a convolution (on a sphere) of

the coefficients of δN_n^m with those from the ocean function (tabulated by Munk and MacDonald, 1960); the result is a leakage of energy from different degrees and orders into others. Those terms where the geoid estimate is highly accurate may be corrupted by neighboring less accurate terms. This leakage is an analogue of that encountered in one dimensional time series analysis where finite data lengths cause leakage of energy from one frequency used in a Fourier transform to another, and where the finite data length prevents resolution of frequencies insufficiently separated. The remedies on the sphere are the same as those in one-dimension and are discussed by Tai (1983). Our chief concerns are to minimize leakage effects and to obtain an estimate of the resolution for coefficients of neighboring spherical harmonics.

GEM-9 and Rapp's (1983) estimate of S were used to produce our first set of results (Fig. 1). To further reduce the leakage, the geoid model was modified. As pointed out before, for geophysical purposes, it often suffices to use $\hat{N} \sim \hat{S}$. Lerch et al. (1982) produced just such a gravity model which they denoted PGS-S4. PGS-S4 is more accurate than GEM-9, because the model error of PGS-S4 is about the size of $|\zeta|$, whereas that in GEM-9 is several times the size of $|\zeta|$. However PGS-S4 cannot be used directly, because it contains ζ. But PGS-S4 <u>can</u> be used to reduce the leakage from large error terms beyond the cutoff degree n_c. So a mixed model was produced for our purposes; it is equal to GEM-9 for terms of degrees less or equal to n_c, and assumes PGS-S4 vaues for terms higher than degree n_c. We write it as $\hat{N}_2 = \hat{N}-\delta\hat{N}_2$. Because $|\delta\hat{N}_2|<|\delta\hat{N}_1|$ for $n>n_c$, $\overline{|\delta\hat{N}_{2n}^m|}<\overline{|\delta\hat{N}_{1n}^m|}$, on average.

The basic procedure was as follows. The difference between S and the gravity model was taken over the oceanic region bounded by 70°N and 70°S latitude. The areal mean was removed and a 10°x10° running average was made

to remove terms of degrees higher than 36. The result was then expanded in terms of spherical harmonics complete to degree and order 36. After some experimenting, the cutoff degree was chosen to be degree $n_c = 6$, thus removing from the result any direct effects of having used S as an estimate of N beyond the cutoff. A second estimate, $\hat{\zeta}_2$, based on this revised procedure is depicted in Fig. 2. It differs in detail from figure 1. The degree variances of ζ_1 (derived from GEM-9) and $\hat{\zeta}_2$ (derived from the mixed model) are listed in Table 1.

3. Results

The altimetric surface S determined by Rapp (1983) represents an estimate from three months of data in the summer and autumn of 1978. We would expect for a variety of reasons, a dynamic height surface based on it to differ from other estimates. Previous estimates have been made by a number of investigators (e.g. Reid and Arthur, 1975 for the Pacific Ocean, and Levitus, 1983 for the global ocean). All these other estimates are based upon time and space averages of hydrographic data. They should differ from our estimate in three primary ways: 1) The hydrographic estimates are averages of data acquired over many years; our estimate is much closer to being a "snapshot" in late 1978; 2) Hydrographic estimates are computed relative to an arbitrarily chosen pressure surface; our estimate is an absolute one in the sense that the reference surface is the geoid; 3) We have removed all wavelengths shorter than about 6,000 km from our estimate whereas the hydrographic estimates contain all wavenumbers, although they are subject to gross and highly variable aliasing as a function of location.

Regarding this third point, note that Roemmich (1983, private communication) has shown that the comparatively large station separation of the IGY Atlantic sections generates a long wavelength alias owing to the

non-resolved mesoscale. The El Niño signal in the tropical Pacific can amount to 30 to 40 cm; with irregular sampling, some of this will appear in the large scale average. The altimetric surface is not entirely free of aliasing, but the sampling is globally uniform and the error is much reduced (see Wunsch and Zlotnicki, 1983).

We may however isolate several features of our result to compare with Levitus' (1983) global chart (figure 3). All the major subtropical gyres in both $\hat{\zeta}_1$ and $\hat{\zeta}_2$ are in the correct sense. The clearest result is achieved in the North Pacific where the spatial scales of flow are maximum and thus most likely to survive the filtering. In contrast with the result of Tai and Wunsch (1983), the filtered flow is more zonal and the center of the gyre is closer to the western boundary. But the low in the eastern North Pacific (Tai and Wunsch, 1983) is still present in Fig. 1. This low is largely eliminated in Fig. 2, implying that it is probably caused by leakage stemming from terms of degrees higher than 6. There are two gyres in the South Pacific, in contrast to Levitus' chart, but consistent with direct surface current observations (Meehl, 1982; but Ekman transports complicate direct comparison). The flow field of the Indian Ocean in Fig. 2 looks more conventional. However the low to the southeast of Africa is virtually unchanged from Fig. 1. If its origin is in geoid error, then these are most likely errors in degree 1-6. Generally speaking, the slopes of the altimetric results are steeper than those derived solely from hydrographic data. Although the steep slope between the North Pacific and the Indian Ocean could be due to orbital error (i.e., the fixed orbit in the crossing arc adjustment process (Rapp, 1983), one expects a nearly instantaneous snapshot (3 months is a very short interval when discussing the general circulation) to have stronger highs and lows than a multiyear average.

4. Discussion

We are aware of at least three other attempts to produce surfaces similar to that shown in figure 1,2. Engelis' (1983) is perhaps closest to ours in methodology, but he fails to discuss the resolution problem and did not smooth the data to reduce aliasing. Low-pass filters other than the spherical harmonic expansion were used in the other two attempts. Douglas, Agreen and Sandwel (1983) confine themselves to an estimate based upon one three day period (the fact they succeed as well as they do with so little data is very encouraging for future, higher precision, satellites than Seasat). Cheney and Marsh (1983, private communication), used PGS-S4 directly. Their results are difficult to interpret, since they have removed a portion of ζ in the difference (3). It is thus impossible to know which of the small scale features their figures display should be taken seriously.

For a slightly more direct comparison, we show in figure 4, the Levitus' (1982) average dynamic topography filtered the same way as Fig. 1 and 2 and to leave only the wavenumber components out to degree and order 6. The comparison with our figure 2 remains quite good in the Pacific Ocean (where we anticipate the best results). But the filtered version of Levitus' surface fails to show the closure of the subtropical gyre-unlike the atlimetric results.

Differences of this sort are easily rationalized for all the reasons previously listed.

The reader may wonder what is the use of the dynamic topography of the ocean obtained only in a low-pass filtered form. The answer to this question is straightforward.

As with any physical phenomenon, we may often choose to isolate the physics in specific wavenumber and frequency bands. Filtering in frequency is a commonplace--e.g. we discuss the mesoscale frequency variability in current meter records after having filtered out longer and shorter periods. We can ask whether our physical models are capable of describing that particular band of frequencies and trying to understand relative amplitude and phase relationships. Such band-passed records may fail to satisfy certain physical constraints (e.g. causality or conservation of total energy) but they are nonetheless extremely useful. Precisely the same applications can be made in wavenumber space: do our general circulation models reproduce the observed long wavelengths in sea surface elevation? If they fail to do so, where, and with what time scales do they fail? Can we force general circulation models of the ocean with low-pass filtered sea surface elevation boundary conditions and have the models compute the shorter wavelengths for us? Wunsch (1983) discusses the general problem of inference from filtered sea surface elevations.

Acknowledgement

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Table 1. Square roots of degree variances, in centimeters, of $\hat{\zeta}_1$, $\hat{\zeta}_2$, and $\hat{\delta N}_1$ (for GEM 9).

Degree	$\hat{\zeta}_1$	$\hat{\zeta}_2$	$\hat{\delta N}_1$
1	43.2	41.8	-
2	28.7	29.4	3.9
3	14.6	14.5	11.8
4	21.7	16.6	9.1
5	19.7	22.1	19.8
6	34.7	30.1	16.3
7	31.5	20.8	29.4
8	33.4	20.8	24.9
9	30.7	20.1	37.9
10	31.0	21.8	34.5

Figure Caption

Fig. 1. Absolute dynamic topography estimate, $\hat{\zeta}_1$, in meters. It is viewed through a low-pass filter corresponding to summing spherical harmonics complete to degree and order 6.

Fig. 2. Same as Fig. 1 except for $\hat{\zeta}_2$.

Fig. 3. Mean annual dynamic topography, in meters, surface relative to 2000 db (from Levitus, 1982).

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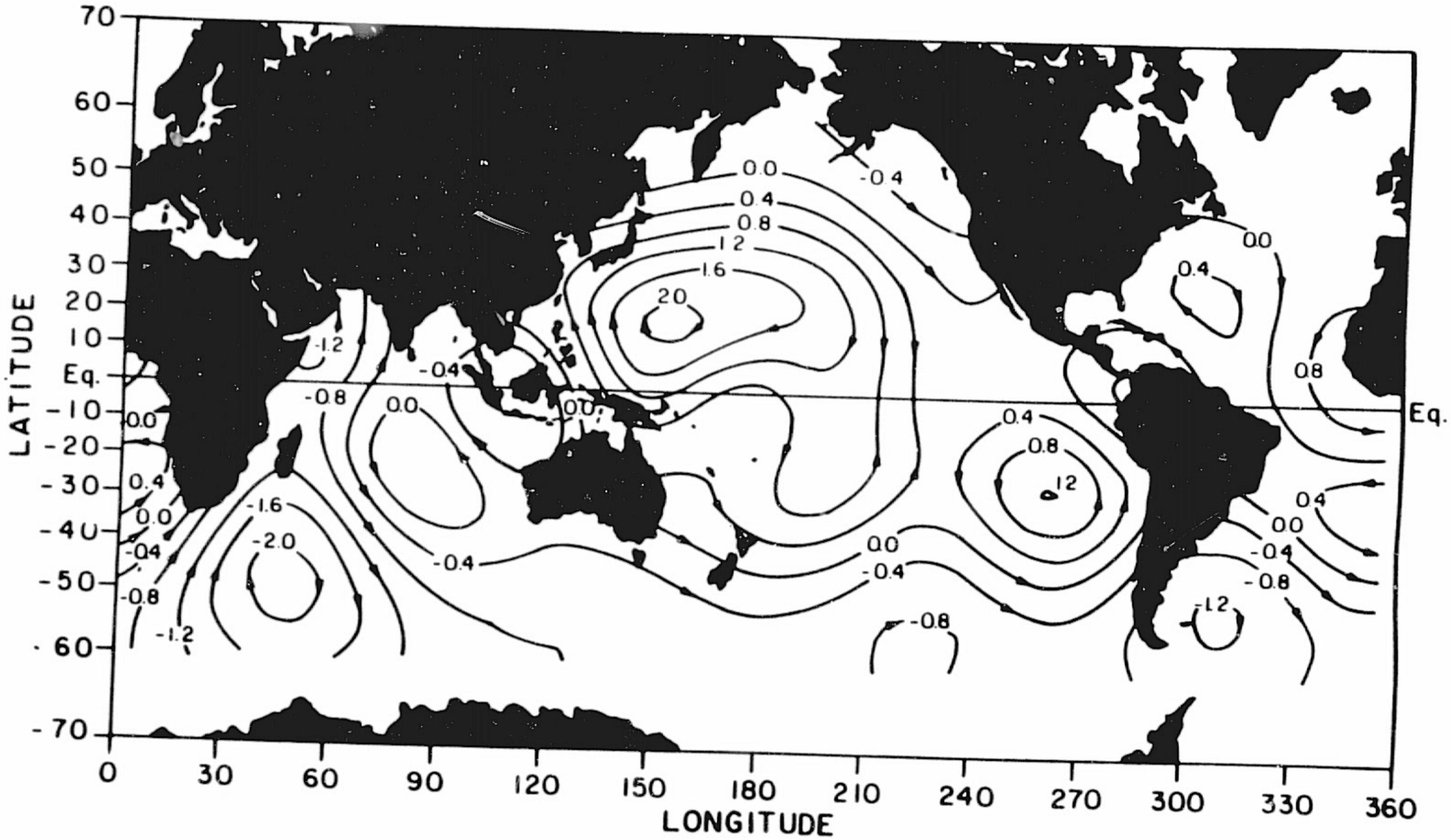
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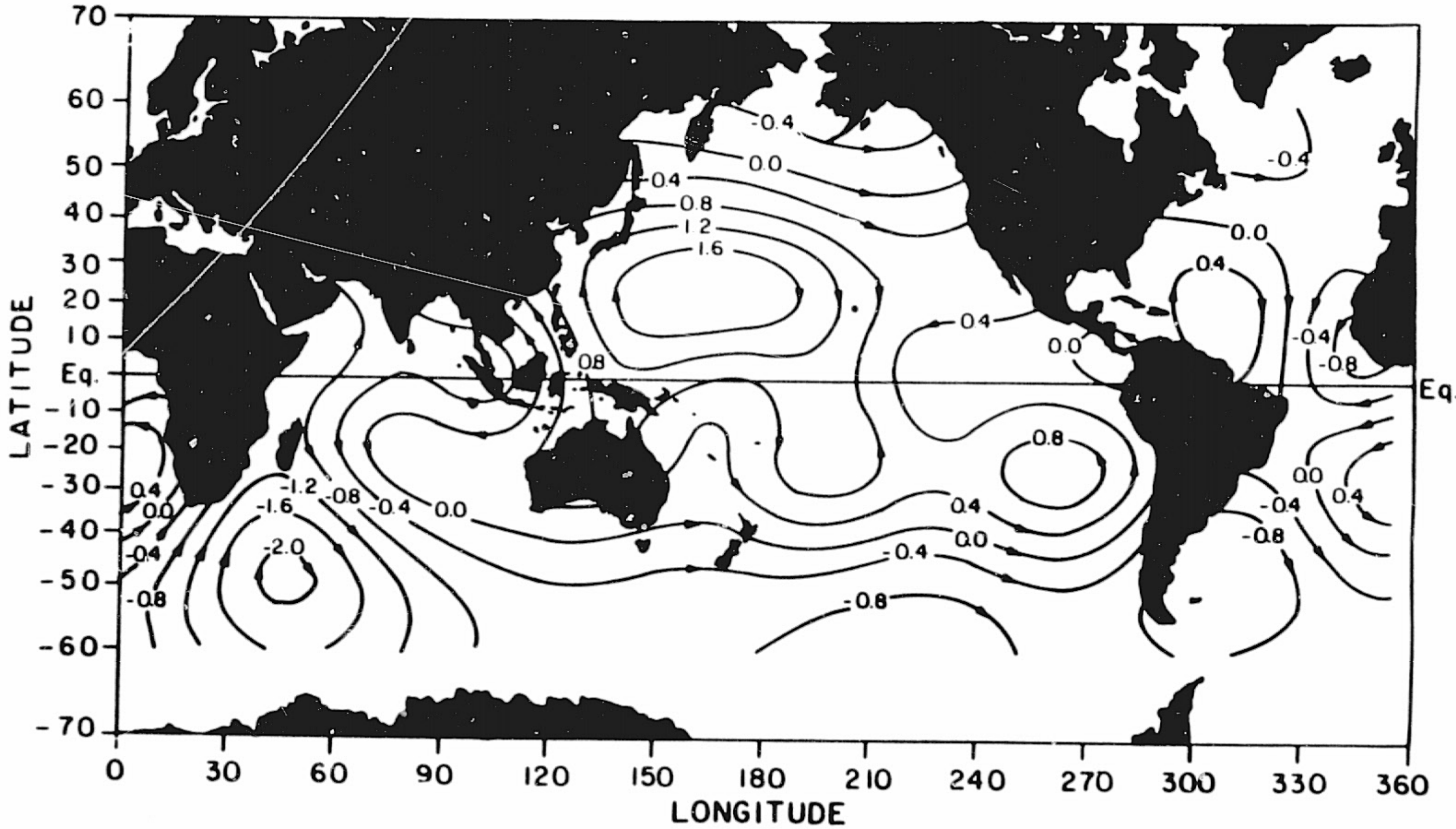

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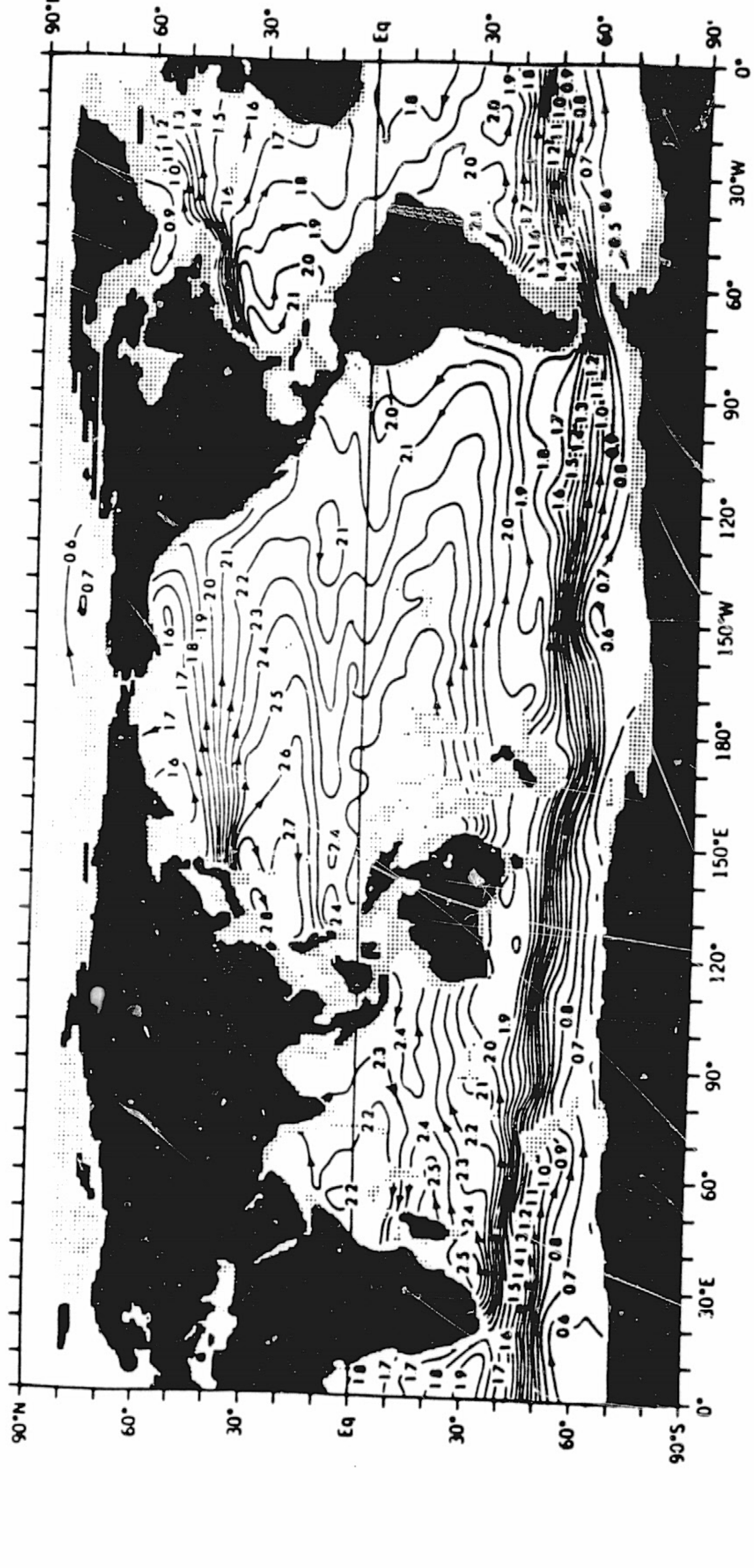
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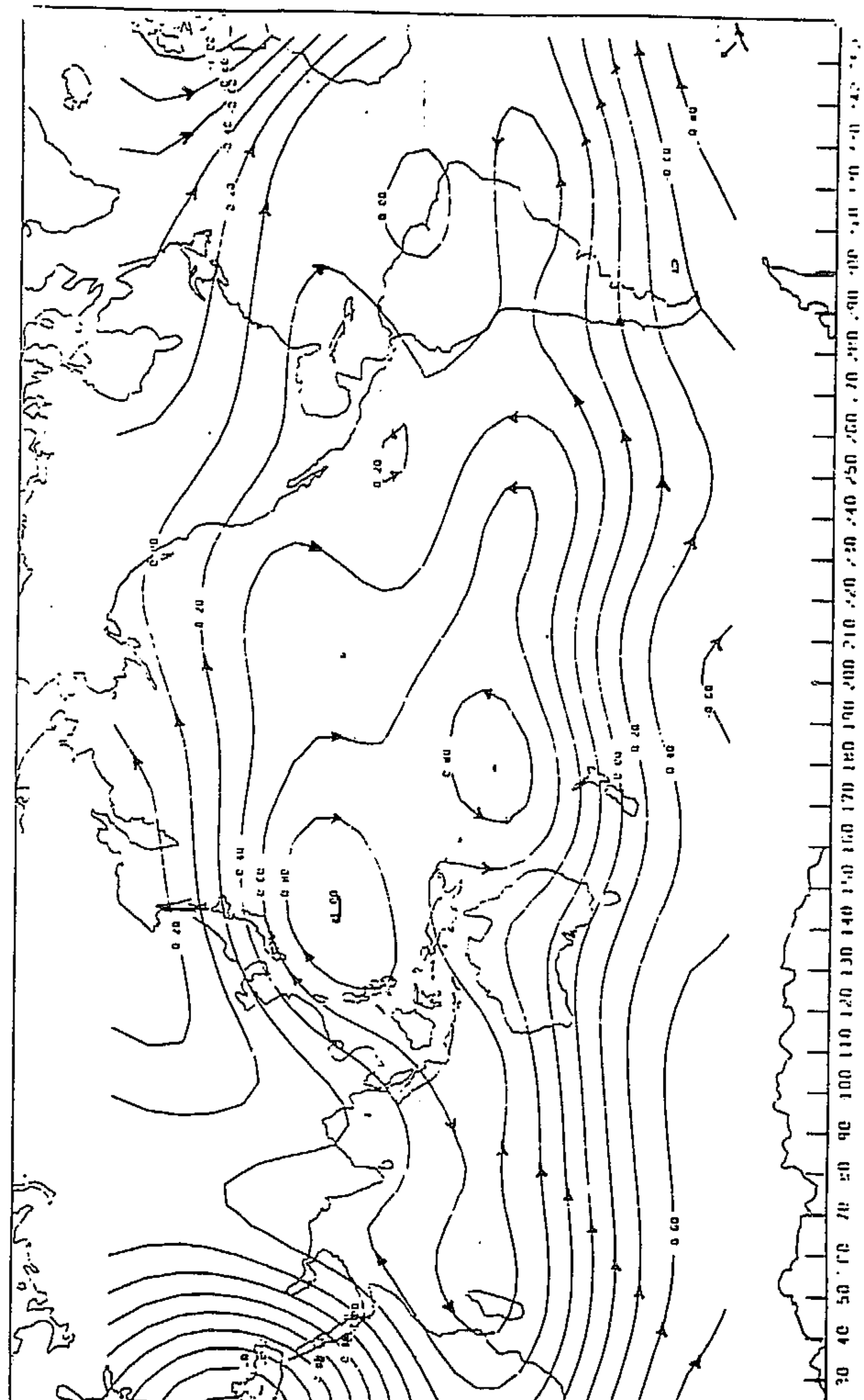

APPENDIX D

On Inferences Concerning the Large Scale Ocean Circulation From Remote and Integrating Measurements

DRAFT

by Carl Wunsch

13 June 1983

1. Introduction

The advent of technologies for observing the ocean on the largest scales has opened up some interesting questions concerning methods for making inferences about the three dimensional ocean circulation and variability. Some of these technologies, e.g. satellite based, or acoustical, are sufficiently different in concept from more conventional methods (e.g. current meters) that the relationships between the observable and the quantities of physical interest are not always obvious. Here we will focus on two novel technologies--ocean acoustic tomography and satellite altimetry--which present different problems of physical inference--and discuss procedures for using the measurements to make statements about the ocean. These two technologies were among the ones discussed by Munk and Wunsch (1982a) as candidates for a system for basin and global observations of the ocean. The exclusion of other methods (e.g. Sofar floats) is not meant to imply that tomography and altimetry are necessarily entire, or even partial, answers to the difficult problem of oceanic observation. But they are specific, and under active study, and they thus provide examples of various possibilities.

The process of making inferences about the ocean from observation is a necessary amalgam of data and knowledge garnered from a vast array of sources--ranging from independent and different observations, to dynamics, to "hunches". In the meteorlogical context, much data analysis is done by the assimilation of observations into large complex, numerical models (see for example, Bengtsson, et al., 1981). In the oceanographic context, studies in this mode of tomography and altimetry are underway, (P. Malanotte-Rizzolli, private communication, 1982). But the models are complicated, and the precise means by which one uses them for assimilation are not yet clear.

Here we will take a different approach. We will demonstrate the joint use of realistically simulated altimetry and tomography with a dynamical model of the ocean circulation. We will emphasize the steady (i.e. time independent) circulation problem, because the linearized time dependent problem is much easier, and its solution was sketched previously by Wunsch and Gaposchkin (1980) and Munk and Wunsch (1982a, Appendix D).

The model we will use (that of Pedlosky, 1969) is very specific and has clear dynamical limitations. In what follows, we will attempt therefore, to do two things--to formulate the problem generally, to demonstrate how one would proceed, almost certainly numerically, with any model; we will also show how to proceed analytically with this particular model as an interesting example of the possibilities in particular cases.

2. A Model

Pedlosky (1969) analyzed the response of a linear, diffusive ocean to an imposed windstress and surface temperature. The basic stratification was taken to be a linear function of the vertical coordinate, z, in order that the lowest order heat equation be satisfied identically. The model was not expected by him to be very realistic; but even this comparatively simple physics leads to a surprisingly intricate solution--as Pedlosky shows. From our point of view, the major virtue of the model is its completeness--one can fully understand the relationship between any dynamical variable and the requirements of all the boundary conditions--including the surface forcing and those on the meridional and zonal walls. In more complex models, e.g. the non-linear thermocline models, there is no explicit relationship available that indicates the structure of the dynamical fields and their relationship to the boundary conditions. The linear model of Rattray and Welander (1976) is another useful candidate for an analysis such as ours; it has a more realistic basic state and a more active connection between the interior and the western boundary current, but a less interesting relationship between the interior vertical velocity and the thermocline structure. In any event, we have chosen the Pedlosky model for our examples. His model and our procedures are meant to be an analogue, or metaphor, for more realistic models and procedures.

We begin with the interior fields. In Pedlosky's notation, (but using a subscript I for "interior" rather than his T), in the ocean of Figure 1, he finds that away from the oceanic sidewall boundary layer we can write

$$u_I(x_o, y_o, \zeta) = \frac{2}{\pi} \int_0^\infty \cos k\zeta \, \frac{1}{fk} \frac{\partial \Theta_I}{\partial x} \, dk \tag{1}$$

$$v_I(x_o, y_o, \zeta) = \frac{2}{\pi} \int_0^\infty \cos k\zeta \left(- \frac{1}{fk} \frac{\partial \Theta_I}{\partial y}\right) dk \tag{2}$$

$$p_I(x_o, y_o, \zeta) = \frac{2}{\pi} \int_0^\infty \cos k\zeta \, \frac{\Theta_I}{k} \, dk \tag{3}$$

$$\theta_I(x_o, y_o, \zeta) = \frac{2}{\pi} \int_0^\infty \sin k\zeta \, \Theta_I(x_o, y_o, k) dk \tag{4}$$

$$\Theta_I(x_o, y_o, k) = C(k)\exp\{- \frac{k^4 f^2}{2\beta} (1-x_o)\} + \int_{x_o}^{1} \frac{f^2}{2\beta} [k^3 T_o(x', y_o) - k(\sigma s)^{1/2} \, \varepsilon_\omega/\varepsilon_T \, \hat{k}\cdot\nabla \times (\frac{\tau}{f})] \exp\{\frac{-f^2}{2\beta} k^4 (x'-x_o)\} dx' \tag{5}$$

$$C(k) = \{ \int_0^1 [1-\exp(\frac{-f^2}{2\beta})] \frac{2\beta}{f} dy\}^{-1} \{\int_0^1 dy' \int_0^1 dx \int_x^1 dx' \frac{f^2}{2\beta} [k^3 T_o(x', y')$$

$$-(\sigma s)^{1/2} \, \varepsilon_w/\varepsilon_T \, \hat{k}\cdot\nabla \times (\underline{\tau}/f)] \exp\{-k^4 f^2/2\beta \, (x'-x)\}$$

$$-(\sigma s)^{1/2} \, \varepsilon_w/\varepsilon_T \int_0^1 dx' [\frac{3}{k^4 + \frac{4\sigma s}{Ef^2}}] \, \tau^{(x)}(x', 1)$$

$$+(\sigma s)^{1/2} \, \varepsilon_w/\varepsilon_T \int_0^1 dx' [\frac{k^3}{k^4 + \frac{4\sigma s}{Ef^2}}] \, \frac{\tau^{(x)}(x', o)}{2f(o)} \tag{6}$$

where β, f and all other variables are non-dimensional, E is an Ekman number, ε_T, ε_W are thermal and mechanical Rossby numbers, T_O and $\underline{\tau}$ are the imposed surface temperatures and windstress, σ is the Prandtl number and S a Burger number. The system has been non-dimensionalized so that $0<x<1$, $0<y<1$ and ζ is the nondimensional depth $-\infty<\zeta<0$ with $\zeta=0$ being the sea surface.

3. Altimetric Measurements

3.1 Perfect Data

We will first consider the case in which an altimetric satellite measures the sea surface topography relative to the geoid. Initially we will assume that the measurement is perfect and only later discuss the effects of errors and sampling gaps. Complete discussions of altimetric systems can be found in Wunsch and Gaposchkin (1980), the TOPEX Science Working Group (1981), Fu (1983), Stewart (1983).

From the model, the surface pressure field of the ocean is (combining 3, 5, 6)

$$p_I(x_o,y_o,o) = 2/\pi \int_0^\infty \frac{C(k)}{k} \exp\left[\frac{-k^4 f^2}{2\beta}(1-x_o)\right]dk$$

$$+ \frac{2}{\pi}\int_0^\infty \frac{f^2}{2\beta}\left[k^2 T_o(x',y_o)-(\sigma s)^{1/2}\frac{\varepsilon_\omega}{\varepsilon_T}\, k\cdot\nabla x\left(\frac{\tau}{f}\right)\right]\exp\left[\frac{-f^2}{2\beta}k^4(x'-x_o)\right]dx' \qquad (7)$$

It is supposed that by use of the altimetry as described in the above papers, we can make a determination of $p_I(x_O, y_O, 0)$ and this is our "data". What inferences can we make about the full three-dimensional ocean circulation from this measurement alone?

The very first step in a problem such as this is to decide what is to be regarded as "known" and with what certainty. For present purposes, we will idealize the situation and suppose that the model physics is complete--i.e. we will discard the (usually realistic) possibility that the model is incapable of accounting for the observations (a simple analogue is the case where we assert that some data is fittable with a straight line, discarding the possibility that a quadratic is really required. We can make tests *a posteriori* to see if a more complex structure should have been included).

The only potentially uncertain parameters of Pedlosky's model then are the physical coefficients, i.e. eddy coefficients (appearing in the guise of Ekman and Prandtl numbers), the gross stratification, etc., and the wind and thermal forcing at the surface appearing through $T(x_O, y_O)$ and $\underline{\tau}$. For illustration, we will regard the Ekman number and other problem parameters as given and known with total certainty (it is easy to extend the treatment to include these parameters).

We are now faced with a linear inverse problem--to make inferences about $T_O(x_O, y_O)$ and $\underline{\tau}$ from measurements of p_I through the vehicle of equation (7). In a linear problem,

there is no necessity to work with perturbations. However most real problems will be non-linear and probably to be approached through a perturbation procedure. Therefore, suppose we have an a priori estimate of both T_O and $\underline{\tau}$ leading to a known a priori estimate of p_I. We thus measure the perturbation

$$\Delta p_I(x_O, y_O, 0) = F(\Delta T_O, \Delta \underline{\tau}) \tag{8}$$

where F is the same operator appearing in eq. 7 but applied instead to the perturbation fields.

Now it has been demonstrated that one can directly measure with scatterometers the wind stress over the sea from space (e.g. Chelton et al., 1981; Satellite Surface Stress Working Group, 1982). But the temperature field occurring in (7) is not the sea surface temperature, it is probably best interpreted as the temperature at the base of the mixed layer. There is no known way of making direct observations of this field from space. So we can pose the following useful question: is the measurement of the surface pressure field by altimetry a useful substitute for the direct determination of the thermal forcing? If it is, then the altimetric measurement would permit computation of the full three dimensional circulation. For illustration purposes then, let us assume that $\underline{\tau}$ is known (this is not a requirement for what follows and in practice one would almost certainly carry both the thermal and wind fields as unknowns, with satellite determined pressure and wind and any other observations as the data in a joint inversion problem).

To be consistent with Pedlosky's underlying assumptions, put $\iint \Delta T_O(x',y')dx'dy' = 0$.

Under these circumstances, (8) reduces to

$$\Delta p_I(x_O,y_O,o) = \frac{f^2}{\pi\beta}\int_{x_O}^{1} dx'\,\Delta T_O(x',y_O)\int_{o}^{\infty} k^2 \exp[-\frac{f^2}{2\beta}k^4(x'-x_O)]dk \tag{9}$$

Performing the integration over k, we have

$$\Delta p_I(x_O,y_O,o) = \frac{\Gamma(3/4)}{4\pi}\left(\frac{f^2}{\beta}\right)^{1/4}\int_{x_O}^{1}\frac{\Delta T_O(x',y_O)dx'}{(x'-x_O)^{3/4}} \tag{10}$$

Eq. (10) is an Abel-Volterra integral equation (see Sneddon, 1972, p. 209) with solution

$$\Delta T_O(x_O,y_O) = \frac{-4\sin\frac{3\pi}{4}}{\Gamma(3/4)(f^2/\beta)^{1/4}}\frac{d}{dx_O}\int_{x_O}^{1}\frac{\Delta p_I(t,y_O,o)}{(t-x_O)^{1/4}}dt \tag{11}$$

Thus complete determination of the thermal perturbation is possible from the pressure perturbation alone; the result supports the idea that even in the absence of thermodynamic information, altimetry plus scatterometry will very highly constrain any plausible model of the ocean circulation.

3.2 More realistic data

Any real altimetric data will contain noise, have some finite spatial coverage, and generally be incomplete. While the expression (11) is a useful theoretical tool, one needs to understand the extent to which inversion for ΔT_O (or for $\Delta\tau$) is practical, and stable. More important, (11) is model dependent; we would prefer a formalism that is not dependent upon the accidents of a particular model choice.

In the general context of inverse theory, we have a set of observations Δp, unknown model function ΔT_j with a known relationship (9 or 10) between observations and model, and some _a priori_ understanding of the noise in the observations. We can proceed to make inferences about ΔT_j in a variety of ways: e.g. the form (10) can be used immediately in the Backus and Gilbert (1967, 1968) inverse formalism (see especially Parker, 1977). Instead, we re-write (9) in discrete form

$$\Delta p_I(x_i, y_i, o) = \sum_j a_{ij} \Delta T_o(x_j, y_j) \qquad (12a)$$

or

$$\Delta p_I = \underset{\sim}{A}\, \underline{\Delta T}_o, \quad \underline{\Delta p}_I = \{\Delta p_I(x_i, y_i)\}, \qquad (12b)$$

$$\underline{\Delta T}_o = \{\Delta T_o(x_j, y_j)\}$$

(x_i, y_i) being observation points and (x_j, y_j) being places where ΔT_o is sought. We could use either the singular value approach of Wunsch (1978) and Munk and Wunsch (1979), or the optimal estimation approach of Liebelt (1967), Bretherton _et al._ (1976), Cornuelle (1983), Zlotnicki (1983). This latter method is particularly useful when we are willing to make assertions concerning the spatial covariances of the data errors and of the temperature perturbations. All of these methods can be shown to be equivalent (see Zlotnicki, 1983, for a review). Wunsch and Zlotnicki (1983) have provided estimates of the expected error covariances in the altimetric

measurement and with the formulas of Liebelt (1967) we can write down explicitly the error with which we can determine ΔT

$$<(\hat{\Delta T}_O - \Delta T_O)^2> = \underset{\sim}{C}_{TT} - \underset{\sim}{C}_{TT}\underset{\sim}{A}^T(\underset{\sim}{A}\underset{\sim}{C}_{TT}\underset{\sim}{A}^T + \underset{\sim}{C}_{NN})^{-1}\underset{\sim}{A}\underset{\sim}{C}^T_{TT} \qquad (13)$$

where $\hat{\Delta T}_O$ is the best (minimum variance) estimate we can make of ΔT_o, C_{TT} is the covariance of the true field ΔT_O, and C_{NN} is the covariance of the errors in the measurement of $\Delta \dot{p}$.

Because (Tai and Wunsch, 1983) in practice the altimetric measurement will define the ocean circulation only on long wavelengths ("long" being defined in terms of the error budgets of geoid estimates), it is particularly important to note that spatially averaged value. of Δp_I are invertible for spatially averaged values of ΔT_O. The model itself provides the "best-estimate" of the unobserved short wavelengths - forced to be compatible with the long wavelength observations and the known dynamics.

3.3 Measuring Short Scales

It is typical of fluid flows that widely different spatial scales are linked both kinematically and dynamically and the same is true in this particular model. Pedlosky shows, for example, that the structure of the western boundary current can be written explicitly in terms of θ_I. If we have managed to construct a regional geoid (and these exist, e.g. Zlotnicki, 1983; Marsh and Chang, 1977) that includes the Gulf Stream system, then an altimetric measurement on the short, O(100 km), scale will constrain and add information to

knowledge of the larger and intermediate scales. It seems unnecessary to display the explicit relationships as they are structurally the same as those already described for the interior scales.

4. Acoustic Tomography Measurements

Munk and Wunsch (1982a,b) discuss the realistic possibilities for velocity tomography, the ability to determine areal average vorticities and to infer areal average vertical velocities. A large number of possibilities exist for using and interpreting these measurements. Here we will examine only a simple example.

From the expressions (2,5,6) we have explicit relationships between the velocities and the forcing fields. We assume following Munk and Wunsch (1982b), that we are able to make estimates from velocity tomography of u,v as a function of depth for fixed x_0, y_0. The surface forcing by temperature and wind is still linearly related to the data--again setting the stage for a linear inverse problem. A linear combination of (1,2), as in a tomographic circulation (vorticity) measurement, remains a linear combination of the unknown fields--preserving the structure of the problem. To reduce the present problem to its simplest possible terms, we suppose once more that we have estimated $\underline{\tau}$ from a scatterometer in space and that our tomographic array is sufficiently large to average out the effects of mesoscale variability, but sufficiently small that we can regard the

measurement as applying to a single point, x_o, y_o, in the ocean. (The great virtue of tomography is its natural integrating behavior).

Consider then the measurement of the perturbation meridional velocity

$$f\Delta v_I = \Delta \frac{\partial p_I}{\partial y_o}(x_o, y_o, \zeta)$$

$$= \frac{2}{\pi}\frac{f^2}{2\beta}\int_0^\infty \cos k\zeta dk \int_{x_o}^1 \Delta \frac{\partial T}{\partial y_o}(x_o' y_o)\exp\{-\frac{k^4 f^2}{\beta}(x'-x_o)\}dx' \quad (14)$$

This expression shows that the meridional velocity is determined by, and thus contains information concerning, the forcing fields only to the east of the observation point and is a consequence of the Sverdrupian interior in Pedlosky's model (a similar result applies to the altimetry). More complex models in which the interior is determined in part by the western boundary currents would have a different information structure.

With adequate vertical resolution from the tomography, we will be able to determine $f\Delta v_I(x_o, y_o, \zeta) = \frac{\partial}{\partial y_o}\Delta p_I(x_o y_o, \zeta)$ and to find their Fourier transforms $f\Delta\hat{v}$, $\frac{\partial}{\partial y_o}\Delta\hat{p}(x_o, y_o, k)$. Thus

$$f\Delta\hat{v}_I(x_o, y_o, k) = \frac{\partial}{\partial y_o}\Delta\hat{p}_I(x_o, y_o, k)$$

$$= \frac{2}{\pi}\frac{f^2}{2\beta}k^2\int_{x_o}^1 \Delta T_o(x', y_o)\exp\{\frac{-k^4 f^2}{\beta}(x'-x_o)\}dx' \quad (15)$$

or

$$\int_{x_o}^{1} \frac{\partial}{\partial y_o} \Delta T(x',y_o)\exp\{\frac{-k^4 f^2}{\beta} x'\}dx' = \frac{\pi\beta}{2fk^2} \exp\{-\frac{k^4 x_o f^2}{\beta}\}\Delta\hat{v}_I(x_o,y_o,k) \tag{16}$$

We thus have an expression for the Laplace transform (the transform variable is f^2k^4/β) of the thermal forcing $\frac{\partial}{\partial y_o} \Delta T_o(x,y_o)h(x-x_o)$, (h is the Heaviside function) expressed in terms of the vertical wavenumber structure of the meridional velocity field. Eq. (16) can thus be easily inverted to give

$$\frac{\partial}{\partial y_o} \Delta T_o(x',y_o) = f \int_{-i\infty+a}^{+i\infty+a} \frac{\pi}{2k^2} \exp\{\frac{-k^4 x_o f^2}{\beta}\}\Delta\hat{v}_I(x_o,y_o,k)e^{\frac{k^4 f^2}{\beta}} d(k^4) \tag{17}$$

4.2 Numerical procedures

As with the altimetry, the inversion formula (17) is a useful analytical tool, but is model dependent and not easily adapted to understanding the effects of noise and finite resolution in the data. But we can write in general,

$$f\Delta\hat{v}_I(x_o,y_o,k) = \sum_j b_{ij} \frac{\partial}{\partial y_o} \Delta T_o(x_j,y_o) \tag{18}$$

and proceed exactly as described above for the altimetric inversion.

5. Some Concluding Remarks

In practice, we would choose to perform a simultaneous inversion for wind field and thermal forcing as well as any other model parameters less than perfectly known--in terms of all available data including the tomography, scatterometry, and altimetry. The principles outlined here are not model dependent although the detailed structure of the answers we obtain is. A major advantage of model dependent inversions is that the solutions we obtain automatically satisfy the appropriate model boundary conditions--which are constraints contributing important information to the structure of the solutions.

Although it may be possible to find more elegant approaches, in using a numerical model the methodology outlined here, based upon a set of analytical relationships, may also be used. If a perturbation expansion is practical, then one can compute, numerically, the set of partial derivatives of, for example

$$\frac{\partial \Delta T(x',y')}{\partial \Delta P(x,y)}$$

or

$$\frac{\partial \Delta T(x',y')}{\partial \Delta V(x,y)}$$

and proceed as before. A related method is to find, numerically, the appropriate Green's functions for the problem.

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